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Attention of all subscribers is called to the following printer's errors in the May 1952 issue of the AMERICAN JOURNAL OF SCIENCE:

Plate 1 of the article by Teng-Chien Yen is misplaced. It should face page 352.

Lines 14 and 15, page 389, are interchanged.

In a few copies only, pages 385 through 400 have been included twice, and pages 353 through 368 are omitted. If you received one of these copies, please return it to us promptly and we will replace it.

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Plates (photographs), *figures* (line drawings), and *footnotes* should each be numbered consecutively through each article, using arabic numerals. If two photographs form one plate, they are figures 1 and 2 of that plate.

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REFERENCE

Lane, B. H., 1935. Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C., 25c.)

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PLIOCENE-PLEISTOCENE BOUNDARY IN CALIFORNIA COAST RANGES *

W. P. WOODRING

ABSTRACT. The Pliocene-Pleistocene boundary in the Cenozoic marine section of California is drawn at a marked faunal discontinuity, at the base of the Santa Barbara formation of the Ventura basin and at the base of the Lomita marl and Timms Point silt of the Los Angeles basin. The boundary so drawn is not opposed by meager evidence based on fossil vertebrates, nor is it known to violate the recommendation of the Eighteenth International Geological Congress.

A period of deformation and erosion is assigned—somewhat arbitrarily as to duration—to the middle third of the Pleistocene. Marine terrace deposits, which locally reach an altitude of at least 1200 feet, are considered of late Pleistocene age.

INTRODUCTION

THE Pliocene-Pleistocene boundary in the Cenozoic marine section of the California Coast Ranges has been a subject of discussion for many years and there still is disagreement concerning it. In the western part of the Ventura basin, in southern California, where 20,000 feet of Pliocene and Pleistocene marine strata are exposed, the boundary has been placed during recent years at three horizons, the lowest and highest of which represent a stratigraphic range of 3,000 feet.

The relations of the marine formations of presumed Pleistocene age to glacial deposits cannot be directly determined. Eaton (1928, pp. 138-139, fig. 3; 1941), indeed, considered the early Pleistocene of California to be preglacial and the late Pleistocene to be late glacial. No unequivocal evidence is now apparent to prove or disprove his opinion that the early Pleistocene is preglacial. There is general agreement that the late Pleistocene is chronologically late glacial. Despite assignments to glacial and interglacial divisions of the Pleistocene by some paleontologists, the age relations of the marine Pleistocene formations to those divisions appear to be indeterminable at the present time by usual stratigraphic

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and paleontologic procedures. According to present estimates, the formations of late Pleistocene age are close to the maximum age limit of radiocarbon dating, some results of which were recently recorded in this JOURNAL (Flint and Deevey, 1951). The youngest, or even perhaps all, of the late Pleistocene formations, however, presumably are within the limit, and they contain suitable material for use of that method of dating.

In view of the uncertainty concerning the age relations of the marine formations to glacial chronology, the Pliocene-Pleistocene boundary is drawn on the basis of the progressive modernization of the local marine faunas, that is, on the basis of the Lyellian principle (but not strictly on Lyellian percentages) of percentage of modern genera and species. That basis, however, has led to different results when applied to the formations under discussion, depending on whether emphasis is placed on faunas as a whole or on the occurrence of particular species that are held to be reliable indicators of age—a fruitful source of dissension among paleontologists.

PLEISTOCENE MARINE FORMATIONS

The age and correlation of the marine formations referred to the late Pliocene and Pleistocene in the present discussion are shown in figure 1. Not all areas of marine late Pliocene are shown in figure 1, and in some of the areas shown part, or all, of the Pleistocene is nonmarine. No exposed marine deposits in California are referred to the middle Pleistocene. Marine deposits of that age, however, are doubtless buried in the deep parts of the Ventura and Los Angeles basins. According to the classification adopted, the middle Pleistocene is represented, at least in the exposed formations of southern California, by a hiatus corresponding to a period of deformation and later deep erosion. The importance of those events was recognized many years ago by Lawson (1893, pp. 157-160; 1894, pp. 269-271), who thought they marked the end of the Tertiary period. The dating now accepted was first pointed out by Eaton (1928, pp. 131-132, fig. 3) and later by other geologists, notably H. Rodney Gale (in Grant and Gale, 1931, pp. 37-38, 44-45, 60-63), Reed (1933, pp. 255-256, 262-263), and Bailey (1935, p. 491; 1943). Bracketing of the deformation and erosion with the middle third of the Pleistocene is, of course, arbitrary in the light of the present knowledge

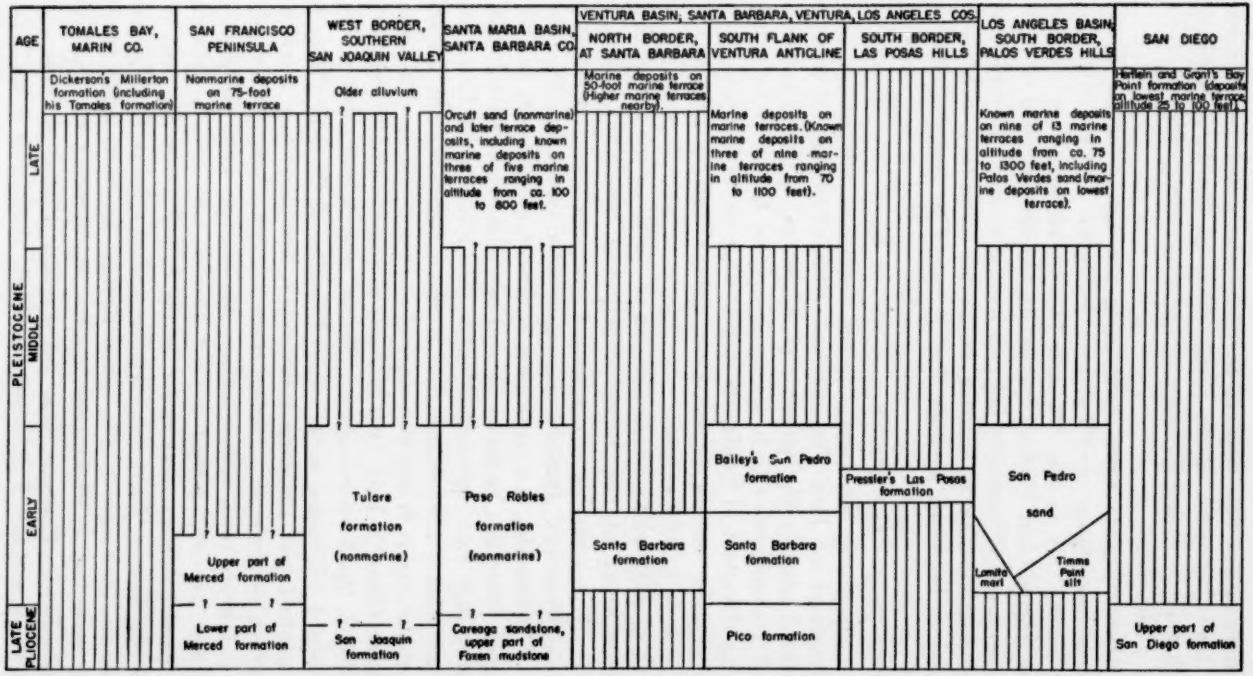
AGE
TOMALES BAY, MARIN CO.
SAN FRANCISCO PENINSULA
WEST BORDER, SOUTHERN SAN JOAQUIN VALLEY
SANTA MARIA BASIN, SANTA BARBARA CO.
VENTURA BASIN, SANTA BARBARA, VENTURA, LOS ANGELES COS.
NORTH BORDER, AT SANTA BARBARA
SOUTH FLANK OF VENTURA ANTICLINE
SOUTH BORDER, LAS POSAS HILLS
LOS ANGELES BASIN; SOUTH BORDER, PALOS VERDES HILLS
SAN DIEGO
PLEISTOCENE
LATE
MIDDLE
EARLY
LATE PLIOCENE
Dickerson's Millerton formation (including his Tomales formation)
Nonmarine deposits on 75-foot marine terrace
Older alluvium
Orcutt sand (nonmarine) and later terrace deposits, including known marine deposits on three of five marine terraces ranging in altitude from ca. 100 to 800 feet.
Marine deposits on 50-foot marine terrace (Higher marine terraces nearby).
Marine deposits on marine terraces. (Known marine deposits on three of nine marine terraces ranging in altitude from 70 to 1100 feet).
Known marine deposits on nine of 13 marine terraces ranging in altitude from ca. 75 to 1300 feet, including Palos Verdes sand (marine deposits on lowest terrace).
Hanlein and Grant's Bay Point formation (deposits on lowest marine terrace, altitude 25 to 100 feet).
Upper part of Merced formation
Lower part of Merced formation
Tulare formation (nonmarine)
San Joaquin formation
Paso Robles formation (nonmarine)
Careaga sandstone, upper part of Foxen mudstone
Santa Barbara formation
Bailey's San Pedro formation
Santa Barbara formation
Pico formation
Pressler's Las Posas formation
San Pedro sand
Lomita marl
Timms Point silt
Upper part of San Diego formation

Fig. 1. Age and correlation of marine late Pliocene and Pleistocene formations of California. Vertical ruling represents hiatus.

concerning the time required for those events and the duration of the Pleistocene epoch itself. Some of the formations of both early and late Pleistocene age may extend into the middle Pleistocene.

Paleontologists agree concerning the age of most of the formations shown as Pleistocene in figure 1. The only disagreement lies in the Pliocene-Pleistocene boundary in the Los Angeles and Ventura basins. In the central part of both basins marine Pleistocene deposits overlie marine Pliocene without discontinuity no matter where the boundary is placed, unless it is placed so high that it would be unacceptable to any paleontologist who has considered the matter during recent years. The boundary shown in figure 1 is drawn to correspond to the most marked faunal discontinuity. The formations of late Pliocene and Pleistocene age shown in figure 1 contain the genera and subgenera, listed in table 1, that are not known to be living, or are not known to be living along the coast of California and farther north in the eastern Pacific. So far as known these genera and subgenera are extinct or locally extinct.

It is quite evident from the list of table 1 that the faunal discontinuity at the Pliocene-Pleistocene boundary shown in figure 1 is a marked discontinuity at generic and subgeneric levels. Moreover, *Anadara* s.s., represented in the late Pliocene by *A. trilineata*, is conspicuous in almost all districts in California where marine formations of late Pliocene age are found. *Lyropecten* s.s., represented in the late Pliocene by *L. cerrosensis*, also is conspicuous, but is less widespread. The faunal discontinuity is fairly well marked at the specific level in some genera found in both late Pliocene and early Pleistocene formations. *Opalia varicostata*, *"Nassa" moraniana* (which, like *Anadara trilineata*, occurs in practically all the late Pliocene of central and southern California), *Strioterebrum martini*, *Mytilus coalingensis*, *Patinopecten healeyi*, *Patinopecten dilleri*, *Ostrea erici*, *Clinocardium meekianum*, and *Platyodon colobus* are well-defined extinct species, not closely related to Recent species, that occur in the late Pliocene but not in the Pleistocene. There are very few such species in the Pleistocene but not also in the late Pliocene. In fact, aside from the three early Pleistocene species representing the genera *Mistostigma*, *Elassum*, and *Adontorhina* (listed in table 1), the early and late Pleistocene *Turritella pedroensis* is the only

species of that class. To be sure, the typical form of the early and late Pleistocene *Calicantharus fortis* and of the early and late Pleistocene *"Cancellaria" tritonidea* are not certainly known to occur in the late Pliocene, but both have close allies of late Pliocene age that may not be of more than subspecific rank.

Paleontologists who place the Pliocene-Pleistocene boundary at a higher level, so as to include in the late Pliocene all, or the lower part, of the Santa Barbara formation of the Ventura basin and the Lomita marl and Timms Point silt of the Los Angeles basin, emphasize the occurrence, in one or more of those formations, of the following species: *Terebratalia hemphilli*, *Crepidula princeps*, *Pecten bellus*, *Pecten stearnsii*, and *Chlamys opuntia*, all of which these paleontologists hold to be reliable indicators of Pliocene age (Bailey, 1935, p. 494; 1943, p. 1562; Grant and Hertlein, 1941; Keen and Bentson, 1944, pp. 11-15). *Arca sisquocensis* and *Ostrea megodon cerrosensis*, both of which are found in the Lomita marl, may be added to that list of species. It is a matter of preference whether these species are to be considered as early Pleistocene survivors or are to be considered as outweighing the faunal discontinuity just described.

Meager evidence based on fragmentary remains of a few fossil land mammals is not opposed to the Pliocene-Pleistocene boundary of figure 1. Teeth identified as *Equus* have been found in the Santa Barbara formation (Bailey, 1943, p. 1562), *Equus* cf. *occidentalis* in Bailey's San Pedro formation of the Ventura basin (Bailey, 1943, p. 1559), and in Pressler's Las Posas formation (Pressler, 1929, pp. 340-341); and teeth identified in the field by Stock as *Equus* cf. *occidentalis* are reported to have been found in the thick section of San Pedro sand on the north flank of the Gaffey anticline at the south border of the Los Angeles basin—strata that evidently include the equivalent of the thick section of Lomita marl on the south flank of the anticline (Woodring, Bramlette, and Kew, 1946, pp. 52, 86).[3]

Practically all paleontologists, invertebrate and vertebrate,

[3] Professor Stock did not carry out his intention of borrowing and studying these fossils. I remember that when he examined them during a trip to the Palos Verdes Hills in 1943 he was disturbed to find, apparently in the early Pleistocene, teeth so similar to those of *Equus occidentalis* (as identified by Merriam) of the late Pleistocene Rancho La Brea asphalt deposits.

TABLE 1

Extinct and locally extinct genera and subgenera in late Pliocene and Pleistocene formations of California

Genus or subgenus	Family	Occurrence	
		Late Pliocene	Pleistocene
EXTINCT			
Echinoid:			
Merriamaster	Scutellidae	San Joaquin formation, Foxen mudstone, San Diego formation.	
Gastropods:			
Mistostigma[1]	Rissoidae(?)		Santa Barbara formation, Lomita marl.
Elassum[1]	Cerithiidae		Lomita marl, Timms Point silt, San Pedro sand.
Apparently unnamed genus or subgenus represented by *"Gyrineum" mediocre*	Cymatiidae	Careaga sandstone, Foxen mudstone, San Diego formation.	
Apparently unnamed genus or subgenus represented by *"Gyrineum" elsmerense*	Cymatiidae	Foxen mudstone.	
Calicantharus	Neptuneidae	Lower part of Merced formation, San Joaquin formation, Careaga sandstone, Foxen mudstone, Pico formation, San Diego formation.	Santa Barbara formation, Bailey's San Pedro formation, Lomita marl, Timms Point silt, San Pedro sand, Palos Verdes sand, Hertlein and Grant's Bay Point formation.
Apparently unnamed genus or subgenus represented by *"Cancellaria" tritonidea*	Cancellariidae	San Joaquin formation, Careaga sandstone, Pico formation.	Bailey's San Pedro formation, San Pedro sand, Palos Verdes sand.
Apparently unnamed genus or subgenus represented by *"Cancellaria" arnoldi*	Cancellariidae	Careaga sandstone, Foxen mudstone, San Diego formation.	
Apparently unnamed genus or subgenus represented by *"Cancellaria" hemphilli*	Cancellariidae	Careaga sandstone, Pico formation, San Diego formation.	
Apparently unnamed genus or subgenus represented by *"Drillia" graciosana*	Turridae	Lower part of Merced formation, Careaga sandstone.	
Pelecypods:			
Lyropecten s.s. (subgenus of *Lyropecten*)	Pectinidae	Careaga sandstone, Pico formation, San Diego formation.	
Adontorhina[1]	Thyasiridae(?)		Lomita marl.
LOCALLY EXTINCT, SURVIVING SOUTH OF CALIFORNIA			

LOCALLY EXTINCT, SURVIVING SOUTH OF CALIFORNIA

Gastropods:			
Trochita	Calyptraeidae	Careaga sandstone, Foxen mudstone, Pico formation.	
Cheilea	Calyptraeidae	Pico formation in Los Angeles.	
Architectonica	Architectonicidae	Careaga sandstone, San Diego formation.	
Pyruclia (subgenus of *Cancellaria*)	Cancellariidae	Careaga sandstone, San Diego formation.	
Pelecypods:			
Arca	Arcidae	Careaga sandstone, Foxen mudstone.	Santa Barbara formation, Lomita marl.
Anadara s.s. (subgenus of *Anadara*)	Arcidae	Lower part of Merced formation, San Joaquin formation, Careaga sandstone, Foxen mudstone, Pico formation, San Diego formation.	
Cunearca (subgenus of *Anadara*)	Arcidae		Palos Verdes sand.
Dosinia	Veneridae	Careaga sandstone, Pico formation, San Diego formation.	Palos Verdes sand, Hertlein and Grant's Bay Point formation.
Apparently unnamed subgenus of *Chione* represented by *Chione picta*[2]	Veneridae		Marine deposits on second terrace in Palos Verdes Hills, Palos Verdes sand.
Lirophora	Veneridae	Careaga sandstone.	
Mexicardia (subgenus of *Trachycardium*)	Cardiidae		Palos Verdes sand, Hertlein and Grant's Bay Point formation.
Varicorbula (subgenus of *Corbula*)	Myacidae	San Joaquin formation, Pico formation.	

LOCALLY EXTINCT, SURVIVING IN NORTHWESTERN PACIFIC

Gastropod:			
Fulgoraria (*Miopleiona* or *Psephaea* of California literature)	Volutidae	Careaga sandstone, Foxen mudstone, San Diego formation.	
Pelecypods:			
Swiftopecten (subgenus of *Chlamys*)	Pectinidae	San Joaquin formation, Careaga sandstone, Foxen mudstone, Pico formation, San Diego formation.	
Pseudocardium	Mactridae	Careaga sandstone.	

[1] These minute genera may be found to be living when the modern fauna at depths of 25 to 100 fathoms is better known.

[2] *Chione picta* was referred to the subgenus *Nioche* by Hertlein and Strong (New York Zool. Soc., Zoologica, vol. 33, pt. 4, p. 187, 1948), but its small size, elongate outline, and interior brown band indicate that it is subgenerically removed from *Nioche*.

agree that the late Pleistocene of figure 1 is indeed to be referred to the late Pleistocene. Remains of land mammals are more numerous in deposits on the youngest late Pleistocene marine coastal terrace, both in marine deposits and in the overlying nonmarine terrace cover, than in the early Pleistocene formations.

The early Pleistocene formations of figure 1 are transgressive and rest with marked unconformity on Miocene or Oligocene formations at the west end of the north and south borders of the Ventura basin, and on early Pliocene or Miocene along the south border of the Los Angeles basin. The marginal deformation and overlap are irrelevant in choosing the series boundary. Their effects, however, have a practical advantage, inasmuch as they agree with the boundary as determined by faunal discontinuity.

TEMPERATURE FACIES OF PLEISTOCENE MARINE FAUNAS

The temperature facies of the late Pliocene and Pleistocene marine faunas has been converted directly into data for estimating isotherm shifts from the time of James Perrin Smith's well-known analysis (Smith, 1919, pp. 134-155) until a recent discussion by Durham (1950, pp. 1257-1260). These faunas cannot properly be used for that purpose unless they have a common denominator of identical, or practically identical, inferred environment. They do not have such a common denominator, if the present environment of species that are still living, or are closely related to species still living, can be trusted as a close guide. In the Palos Verdes Hills (at the south border of the Los Angeles basin), the Lomita marl, Timms Point silt, San Pedro sand, and Palos Verdes sand contain 150 to 350 species of mollusks, a total of about 500 species. The faunal facies of these formations, which have been assigned to glacial and interglacial subdivisions (Gale, in Grant and Gale, 1931, pp. 68-76), has been recently documented and discussed (Woodring, Bramlette, and Kew, 1946, pp. 86-103). The early Pleistocene Lomita marl and San Pedro sand, and particularly the Timms Point silt, contain cool-water species. They are found, however, in associations inferred to represent different depths ranging from 25 to 100 fathoms. No shallow-water northern fauna is known to occur in the Los Angeles or Ventura basin.

The late Pleistocene Palos Verdes sand (the marine deposits on the lowest terrace) has a shallow-water fauna of 250 species, about 20 of which are north of their present range, some species hundreds of miles. The marine deposits on the next older terrace contain seven such species out of a total of 110; those on the fourth terrace (in upward sequence) one southern species out of 90. Other terrace deposits in the Palos Verdes Hills have so far yielded only a relatively meager rock-cliff and tide-pool association that is indistinguishable no matter on what terrace it is found, except that the fourth terrace has yielded a southern species in an association of that character. Though the fossils from at least the lowest terraces may be interglacial or interstadial, they do not represent a Lower California fauna. Morover, the southern species are found in terrace deposits from San Diego northward to Santa Monica, but not in apparently contemporaneous terrace deposits in the Ventura basin and farther north. These relations indicate that the western end of the Santa Monica Mountains marked the boundary between two well-defined late Pleistocene faunal provinces that are unrecognizable at the present time.

At localities in Santa Barbara County north of the Ventura basin terrace deposits contain a fauna like the modern fauna at the same latitude. Contrary to expectation, Dickerson's Millerton formation (Dickerson, 1922, pp. 559-570; Mason, 1934, pp. 85-87, 104-105; Weaver, 1949, pp. 99-104), immediately north of San Francisco Bay, contains 40 species, 14 of which are somewhat north of their present range. As suggested by Dickerson, that formation may be interglacial, or interstadial. According to Mason (1934, pp. 99-106), however, the associated flora has a northern aspect.

RECOMMENDATION OF EIGHTEENTH INTERNATIONAL GEOLOGICAL CONGRESS

The Council of the Eighteenth International Geological Congress unanimously accepted a recommendation, presented by a Temporary Commission, that "the lower Pleistocene should include as its basal member in the type area [Italy] the Calabrian formation (marine) together with its terrestrial (continental) equivalent the Villafranchian" (Anonymous, 1950). Moore (1949) interpreted this recommendation to

mean that the San Joaquin formation,[4] shown in figure 1, is early Pleistocene, on the grounds that it contains a vertebrate fauna of vertebrate paleontologists' Blancan stage and that the Blancan is correlated with the Villafranchian. The meager vertebrate fauna of the San Joaquin formation is considered by vertebrate paleontologists of Blancan age on the basis of the occurrence of teeth described by Merriam as *Pliohippus proversus*, a species now referred to *Plesippus* or *Equus*. There is no agreement, however, among vertebrate paleontologists that the Blancan is to be correlated with the Villafranchian. The Blancan has, in fact, been correlated with the Villafranchian (McGrew, 1944, pp. 42-48; Schultz and Stout, 1948, p. 571), and it has been claimed that the Blanco formation and correlated deposits are late Nebraskan and early Aftonian (Frye, Swineford, and Leonard, 1948, pp. 519-521, fig. 3). The Blancan, however, has also been correlated with the Astian of Italy (Wood and others, 1941, plate 1), which underlies the Villafranchian, and with both Astian and Villafranchian (Simpson, 1947, p. 623).

Until there is agreement that the Blancan is of Villafranchian age, California geologists need not be concerned that they are flouting international agreement by continuing to refer the San Joaquin formation to the late Pliocene. It is doubtful whether any paleontologist familiar with the succession of Cenozoic marine molluscan faunas of western North America would accept the San Joaquin and correlated formations as Pleistocene for any reason other than to encourage international agreement.

[4] Moore's representation, in his figure 1, of the Etchegoin, San Joaquin, and Tulare formations is misleading. At no place where the three formations are exposed is the San Joaquin formation neatly separated from the Etchegoin by a discontinuity, except a minor scour discontinuity similar to numerous discontinuities in both formations. Nor is the Tulare formation neatly separated from the San Joaquin by a discontinuity. In the Kettleman Hills the lower part of the Tulare formation, shown as late early Pleistocene in Moore's diagram, contains an extensive fauna of nonmarine mollusks unlike any known Pleistocene fauna. It is quite certainly not later than late Pliocene (Woodring, Stewart, and Richards, 1941, pp. 103-104).

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ABSOLUTE AGES OF RADIOACTIVE MINERALS FROM THE APPALACHIAN REGION

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ABSTRACT. All available absolute age determinations made on radioactive minerals from the Appalachian region have been reestimated in the light of presently accepted formulae and constants. They appear to record four orogenies. Two, roughly 800 and 600 million years ago, were Precambrian and are recorded only along the western edge of the crystalline Appalachians. The Taconian, 350 million years ago, affected a belt in the western part of the crystalline Appalachians and may also have been represented in eastern Massachusetts. The Acadian, 260 million years ago, affected eastern New England. Curiously the Appalachian orogeny is at best feebly recorded. The igneous rocks of the late Triassic appear to be about 170 million years old.

INTRODUCTION

IN the course of preparing a discussion of the structural history of East Tennessee (part of a report to accompany a new geologic map of East Tennessee, to be published by the Tennessee Division of Geology), I became interested in the extant absolute age determinations made on minerals from the southern Appalachians. With the kind help and encouragement of Professor Adolph Knopf of Yale University, I returned to the original data and undertook to review them in the light of currently accepted formulae and constants. The task was greatly simplified by use of the graphs recently prepared by Wickman (1939, figs. 1, 2, 3; 1944, plate 1), which embody these formulae and constants and enable one to dispense with many tedious calculations. Once launched on the subject, I extended my survey to the central and northern Appalachians as well, and the results are presented herewith. I hope to have listed all the reliable analyses and determinations from which ages can be calculated that have been made on minerals from the Appalachian region.

Several different methods for determining the absolute age of radioactive materials have been worked out, but the fundamental one is still the lead-ratio method, which depends on the ratio of the lead produced by the main radioactive series (uranium-238, actinium or uranium-235, and thorium series) to the parent radioactive substances remaining. Other methods are based on the helium produced by the same radio-

active substances, and on the strontium produced by the radioactive decay of rubidium, but for the Appalachians these others are at present only of corroborative value. Where quoted in this article, ages determined by these other methods are taken directly from the original publication.

Determinations by the lead method likewise differ in reliability. Knopf (1949, p. 4) has divided them into four classes: "Class I, in which U, Th, and Pb have been determined by standard gravimetric chemical analyses and the isotopic composition of the lead has been determined by means of the mass spectrograph; Class II, in which U, Th, and Pb have been determined by standard gravimetric methods and the atomic weight of the lead has been determined chemically; Class III, in which U, Th, and Pb have been determined by standard gravimetric methods; and Class IV, in which a few milligrams of material have been analyzed for their U, Th, and Pb content by microanalytic methods."

There are three Class I determinations on Appalachian minerals; two on two samples from one locality and one on a sample from a different locality and, as it appears, age. As the nature of the data from these determinations permits a double check (triple if Th is present), they are considered of maximum reliability. In addition there are about 50 Class III and IV analyses, age determinations from which involve the assumption that all the lead present is the product of radioactive decay—an assumption justifiable in certain cases but dangerous in others. The procedure used to obtain absolute ages from these kinds of analyses is demonstrated below in the discussion of the Spinelli quarry samarskite.

The helpful advice and critical interest of Professor Knopf are gratefully acknowledged; without them this paper could not have been satisfactorily prepared. The manuscript has been critically read by Mr. Knopf, and by Chester R. Longwell, Philip B. King, Jarvis B. Hadley, John P. Marble, and Marland P. Billings, and I wish to thank them all for their valuable help. Others kindly helped by answering questions about areas with which they are especially familiar; they are cited in the appropriate places in the paper. Shirley P. Glaser drew the figure.

LATER PALEOZOIC DATES

Age determinations.—One of the Class I determinations of absolute age in the Appalachians is on samarskite from

the Spinelli pegmatite quarry in Glastonbury, Connecticut. Chemical analysis of this mineral, by R. C. Wells (1937, p. 115: for mean see Lane, 1935, p. 76; 1936, p. 67), recalculated from oxide to element, gave:

	Pb	Th	U	y[1]	x[1]	Preliminary age[1] (million years)
I	0.278	3.18	7.06	0.0271	0.689	250
II	0.336 0.316	2.92	6.79	0.0346 0.0325	0.699	300 290
Best mean	0.311	3.05	6.91	0.0312	0.694	280

The isotopic constitution of the lead, determined by Nier and others (1941, p. 113) is:

Pb^{204}	Pb^{206}	Pb^{207}	Pb^{208}
0.167	100	7.60	21.3

As calculated from these data, the (chemical) atomic weight is 206. 385. If it is assumed with Nier (p. 114) that common lead has the composition 1 : 18.5 : 15.4 : 38.2[2], there is 9.5 percent common lead in the sample. After common lead has been subtracted from each isotope, the product-parent ratio can be calculated for each radiogenic isotope of lead (it is assumed that $U^{238}/U^{235} = 139$; see Nier, 1939a). When these ratios are substituted in the formulae developed by Keevil (1939, equations 7a, 8a, 9a, p. 199[3]), three separate determinations of age are obtained:

Pb^{206*}/U^{238} .....251 million years,
Pb^{207*}/U^{235} .....255 million years,
$Pb^{208*}/$ Th .....264 million years, where the asterisk indicates that only the radiogenic fraction of the isotope is considered. (These ages can also be read directly from the graphs prepared by Wickman, 1939, figs. 1, 2.)

[1] $y = \frac{Pb}{Th + U}$, $x = \frac{U}{Th + U}$; these are ordinate and abscissa for the graph prepared by Wickman (1944, plate 1) from which the figures in the last column have been read. If, as indicated by the isotopic constitution, there is 9.5 percent common lead, the figures for the mean become Pb – 0.282, y – 0.0283, age 250 million years.

[2] Holmes (1947) assumes a slightly different composition for the common lead, but it makes only a minor difference in the results.

[3] The numerical coefficients in Keevil's equations have been altered to conform to more recent data. The first set of coefficients depends on the disintegration constants for U^{238}, U^{235}, and Th^{232}; the figures here used are as follows:

As Nier (1939b) has pointed out, the age can also be determined from the ratio Pb^{207*}/Pb^{206*}; for this another graph of Wickman's (1939, fig. 3) is available. The present data give a figure of 280 million years, but it should be noted that this determination is not strictly independent of the first two determinations above. It should also be noted that the last digit in all these determinations is very weak; in general, absolute ages by the lead-ratio method should not be quoted beyond the second significant figure.

The age of the Spinelli quarry samarskite may therefore be taken as about 260 million years.

The Spinelli quarry is one of a group of pegmatite quarries east of Middletown, Conn., and a number of chemical analyses of radioactive minerals from these quarries have been made. They are listed below with the age as determined from the graph (Wickman, 1944, plate 1):

	Class	Pb	Th	U	y	x	Age (million years)
Monazite from Hale quarry, Portland, Conn. (Fenner, 1932, p. 330):							
	III	0.1007	7.489	0	0.0135	0	300
Uraninite from Hale quarry, Portland, Conn. (Hillebrand, 1891, pp. 62-63):							
	III	2.91	8.41**	70.52	0.0369	0.893	280
		3.01	8.60**	70.59	0.0380	0.891	290
		2.85	9.06**	70.80	0.0357	0.887	270
		3.03	8.61	72.68	0.0373	0.894	280
		2.86	8.80	72.02	0.0354	0.891	270
Uraninite from Strickland (Collins Hill) quarry, Portland, Conn. (Hecht and Kroupa, 1936, p. 87; for layers of a single crystal):							
(outside)	IV	3.35	2.86	80.14	0.0404	0.966	290
		3.20	3.17	80.20	0.0384	0.962	280
		3.12	3.26	78.86	0.0380	0.960	280
(core)		3.12	3.19	79.00	0.0380	0.961	280

[3](continued)

Nuclide	Disintegration constant (λ)	First coefficient	Reference
U^{238}	1.535×10^{-10}	15.00×10^{9}	Nier and others, 1941
U^{235}	9.80×10^{-10}	2.35×10^{9}	Fleming, Ghiorso, and Cunningham, 1951
Th^{232}	0.499×10^{-10}	46.2×10^{9}	Kovarik and Adams, 1938

The second set of coefficients expresses the ratio of the atomic weights of the parent and product elements. Apparently Keevil (and Holmes, 1947) used the ratio of the atomic weights of the individual isotopes involved. But the mass spectrograph determines isotopic constitution not in weight percent but in atomic percent; hence the coefficient should be the ratio of the atomic weights of the whole mass of the elements in the sample for which the isotopes were determined. The atomic weights of U and Th are here taken as 238.07 and 232.12, in accordance with the latest figures issued by the International Union of Chemistry. The atomic weight for the lead in each mineral is taken as calculated from the isotopic constitution. As a result of these changes, the ages here calculated differ from those previously calculated from the same data, but the differences are mostly in the third significant figure and are therefore unimportant.

(Hecht and Reich-Rohrwig, 1933, p. 287):						
	IV	3.39	3.34	78.81	0.0413	0.959	300
		3.66	2.28	78.82	0.0450	0.972	320
		3.37	3.32	78.71	0.0410	0.960	300
(Føyn, 1938, p. 18):						
	III	2.91	2.88	74.39	0.0377	0.963	270
Uraninite from Rock Quarry Landing, Haddam (Neck), Conn. (Hecht and Kroupa, 1936, p. 89):						
	IV	3.09	1.48	76.60	0.0396	0.981	280

** In these analyses a certain amount of the rare earth elements was included with the thorium. The result would be to make the age as calculated a little low.

The beautiful consistency of these results shows that the whole Middletown pegmatite area is of one age, best dated as 260 million years old by the triply-checked Class I determination on the Spinelli quarry samarskite.

There are two other chemical analyses that may belong to the same general age range in New England:

	Class	Pb	Th	U	y	x	Age (million years)
Uraninite from Ruggles mine, Grafton, N. H. (Shaub, 1938, p. 338):							
	IV	3.37	0.38	76.38	0.0439	0.995	310
Samarskite from Topsham, Maine (Lane, 1937, p. 60):							
	III	0.47(3)	2.66	14.31	0.0279	0.843	220

A Class IV analysis of monazite from Topsham, discussed below, yields, however, a much higher age.

Age determinations by other methods of minerals that appear to belong in this range are as follows:

Lead ratio by alpha count and spectrographic analysis for lead (Larsen and others, 1949, p. 27); 2 determinations on zircon from Conway, N. H., granite:

181 million years
262 " "

Strontium-rubidium ratio in lepidolite (Ahrens, 1949, pp. 251-252):

Newry, Me.	300 million years	
Norway, Me.	280 " "	
Topsham, Me.	200 " "	
Mt. Mica, Paris, Me.	240 " "	
Buckfield, Me.	300 " "	
Auburn, Me.	540 " "	(rejected by Ahrens, p. 255)
Haddam (Neck), Conn.	270 " "	
Middletown, Conn.	450 " "	(rejected by Ahrens, p. 255)
Portland, Conn.	540 " "	" " "

Helium ratio (Hurley and Goodman, 1943, p. 309):

Magnetite from Chester, Mass.	225 million years	(amount of helium too small to be accurately measured, p. 311)
Pyrrhotite from Ducktown, Tenn.	215 " "	
	245 " "	

Geologic setting.—The pegmatites of the Middletown, Conn., area are supposed to be related to the Monson (granodiorite) gneiss (Foye, 1949, pp. 51-52), but Stugard (in preparation), who has worked most recently in the area, is skeptical of the

evidence for the relation. Geologically the age of neither pegmatite nor gneiss is known, except that they are pre-Triassic. The pegmatite at the Ruggles mine, Grafton, N. H., is supposed to be related to the Bethlehem gneiss, a member of the New Hampshire magma series, which is synorogenic to the Acadian orogeny in New Hampshire. The New Hampshire and Connecticut areas are nearly 150 miles apart but not far off strike with each other. About midway between, however, occurs the Greenwich, Mass., allanite, discussed below, which appears to be considerably older.

The Topsham, Maine, area is about 100 miles across the strike east of Grafton, N. H., in rocks of doubtful but probably Paleozoic age (Fisher, 1941, plate 1).

The Conway, N. H., granite lies between Grafton and Topsham, nearer Grafton. It is part of the post-orogenic White Mountain magma series. The Maine localities listed by Ahrens, except Topsham and Auburn, are fairly close together northeast of Conway, but the age relations of the rocks have not been established.

Chester, Mass., is west of the Connecticut River, in a different and supposedly older structural province from Middletown and Grafton. The magnetite came from the well known emery deposit, which Emerson (1898, pp. 143-147; 1917, pp. 159-161) thought was metamorphic, so that the magnetite would be roughly contemporaneous with the metamorphism of the enclosing schists, supposedly early Paleozoic. It is difficult to make the helium age of 225 million years square with this interpretation, but both age and interpretation are subject to question.

The pyrrhotite at Ducktown, Tenn., occurs in sulphide pods in Precambrian or very low Cambrian rocks, but the sulphides are practically post-orogenic and are supposed to have been introduced near the end of the Paleozoic (Emmons and Laney, 1926, p. 64; Ross, 1935, pp. 23-24).

The bulk of the evidence associates the minerals giving ages of 260 ± 50 million years with the Acadian orogeny which gripped eastern New England in the later Paleozoic. This orogeny is usually supposed to have occurred at the end of the Devonian on the basis of evidence in New Brunswick and Gaspé, but even there the evidence suggests rather several pulses scattered through Devonian, Mississippian, and

early Pennsylvanian time (Alcock, 1947). In view of determinations of about 230 million years on Lower Permian minerals from Bohemia and Norway (Knopf, 1949, p. 5), the figure of 260 million years for the Spinelli quarry samarskite favors a later date than late Devonian. Actually the oldest dated beds not affected by the so-called Acadian deformation in central New England, the Rhode Island Coal Measures, appear to be of Allegheny or at most late Pottsville age (Knox, 1944)—about Medial Pennsylvanian.[4] The stratigraphic evidence west of New England indicates two disturbances: one, represented by the Catskill-Pocono detrital wedge, culminating in late Devonian and earliest Mississippian time, the other, represented especially by the Pottsville conglomerate, culminating in early Pennsylvanian time; of the two detrital wedges, the later is coarser and more widespread but considerably thinner, and it overlies a marked disconformity. Thus in central New England the strongest pulse of the Acadian orogeny may well have been as late as early Pennsylvanian. In the Maritime Provinces—Acadia—however, the strongest pulse occurred during the Medial and/or Late Devonian.

It is remarkable that no "Appalachian" date is recorded except by the not entirely satisfactory helium determinations from Ducktown, Tenn., and perhaps Chester, Mass., and perhaps also by the lead-ratio and strontium-rubidium ratio ages from Topsham, Maine. All the published lead-ratio ages in the southern and central Appalachians are early Paleozoic or older.

EARLY PALEOZOIC DATES

Age determinations.—Class I determinations have been made on two samples of uranoan zircon (cyrtolite) from Bedford, Westchester County, New York. The data are as follows:

I. Chemical analysis (Muench, 1931, p. 356):

Pb	Th	U	y	x	Preliminary age (million years)
0.374	tr	7.29	0.0513	1	360

If account be taken of 6.2 percent common lead, the figures become:

0.351	tr	7.29	0.0481	1	340

[4] The age of the Worcester phyllite, which was apparently metamorphosed during this deformation, remains uncertain. David White (1912; quoted by Emerson, 1917, pp. 63-64) concluded that it was certainly Carboniferous, possibly Pottsville. If, however, it were Mississippian, it might have preceded the main deformation in this region.

Isotopic constitution of lead (Nier, 1939b, p. 158):

Pb^{204}	Pb^{206}	Pb^{207}	Pb^{208}
0.095	100	6.89	4.29

indicating 6.2 percent common lead.

Atomic weight of lead (calculated from isotopic constitution): 206.140.

Age from ratios: Pb^{206*}/U^{238} ..... 334 million years
Pb^{207*}/U^{235} ..... 347 " "
(Pb^{207*}/Pb^{206*} ..... 420 " ")

II. Chemical analysis (Muench, 1934):

Pb	Th	U	y	x	Preliminary age (million years)
0.351	0	6.73	0.0522	1	360

If account be taken of 3.0 percent common lead, the figures become:

0.341	0	6.73	0.0507	1	350

Isotopic constitution of lead (Nier, 1939b, p. 158):

Pb^{204}	Pb^{206}	Pb^{207}	Pb^{208}
0.0443	100	6.07	2.22

indicating 3.0 percent common lead.

Atomic weight of lead (calculated from isotopic constitution): 206.099.

Age from ratios: Pb^{206*}/U^{238} .....352 million years
Pb^{207*}/U^{235} .....358 " "
(Pb^{207*}/Pb^{206*} .....385 " ")

The Bedford zircon may therefore be considered to have an age of about 350 million years.

Several analyses of uraninite from Branchville, Conn., about 10 miles east of Bedford, N. Y., give comparable ages:

	Class	Pb	Th	U	y	x	Age (million years)
(Hillebrand, 1891, p. 64):							
	III	4.04	6.33**	74.77	0.0498	0.922	370
		4.03	6.09	75.00	0.0497	0.925	360
		4.04	5.73**	74.26	0.0505	0.928	370
(Boltwood, 1907, p. 79):							
	III	3.5	n.d.	66	0.053	1–	370
(Comstock, 1880, p. 220):							
	III	4.01	n.d.	81.67	0.0491	1–	340
		3.94	n.d.	81.33	0.0484	1–	340

A number of analyses of uraninite from Spruce Pine, Mitchell County, North Carolina, also fall in this age range:

(Alter and McColley, 1942; on three layers of a single crystal, specific locality not given):							
(outside)	III	3.19	0.03	67.65	0.0471	1.000	330
		3.40	0.14	69.92	0.0485	0.998	340
(core)		3.50	0.04	69.65	0.0502	0.999	350
(Føyn, 1938, p. 18; Deer Park mine):							
	III	3.64	2.57	74.20	0.0474	0.967	340
(Boltwood, 1907, p. 79; Flat Rock mine):							
	III	3.3	n.d.	67	0.049	1–	340

** In these analyses a certain amount of the rare earth elements was included with the thorium. The result would be to make the age as calculated a little low.

(Hillebrand, 1891, pp. 65-67; on a single specimen from Flat Rock mine, before and after leaching with HCl):

(before)	III	3.90	2.44	76.96	0.0491	0.969	350
(after)		4.20	2.67**	77.77	0.0522	0.967	370

(von Foullon, 1883, p. 7; "Stony Point"):

	III(old)	3.55	n.d.	81.09	0.0438	1–	310

A monazite from the Deer Park mine, Spruce Pine, discussed below, gives, however, a distinctly older age.

Scattered age determinations in the same range may be added:

Class	Pb	Th	U	y	x	Age (million years)
Uraninite from Marietta, S. C. (Hillebrand, 1892, p. 23):						
III	3.32	1.45	71.19	0.0457	0.980	320
Allanite from Greenwich, Mass. (now under Quabbin Reservoir) (Marble, 1950):						
III	0.032	1.530	0.095	0.0197	0.058	390
Uraninite from Fitchburg, Mass. (Hecht and Kroupa, 1936, p. 88):						
IV	2.52	3.39	50.20	0.0470	0.937	340
Allanite from Blueberry Mountain, Woburn, Mass. (Lane, 1928, p. 6; Holmes, 1931, p. 343):						
III	0.03(6)	2.013	0.11	0.0170	0.052	340
(Hecht and Kroupa, 1936, p. 100):						
IV	0.034	1.76 1.63	0.039	0.0189 0.0204	0.022 0.023	400 430
(Lane, 1939, p. 106):						
(rim) IV	0.0358	1.893	0.0265	0.0186	0.014	400
(core)	0.0311	2.008	0.0099	0.0154	0.005	340

and by other methods:

Impure magnetite from Salem, Mass., gabbrodiorite, helium determination (Hurley and Goodman, 1943, p. 310): more than 330 million years.

Zircon from Statesville, N. C., determined by alpha count and spectrographic analysis for lead (Larsen and others, 1949, p. 27): 355 million years.

Although the incomplete von Foullon analysis from Spruce Pine overlaps the analysis from the Ruggles mine, Grafton, N. H., in general this group of analyses appears to be distinctly older than the Acadian group.

Geologic setting.—The Bedford, N. Y., and Branchville, Conn., pegmatites are supposed to be related to the Thomaston granite of western Connecticut, whose age is geologically unknown other than that it is pre-Triassic. The localities lie

** In these analyses a certain amount of the rare earth elements was included with the thorium. The result would be to make the age as calculated a little low.

a little east of the certainly Precambrian core of the Hudson Highlands in schists of uncertain age, perhaps Precambrian, perhaps early Paleozoic. On the basis of these radioactive determinations, the granite has been dated as Taconian; the minerals would certainly seem to be post-Cambrian as the fossiliferous and uraniferous Upper Cambrian Kolm shale of Sweden is well dated as approximately 440 million years old (Knopf, 1949, p. 5). The existence of a late Ordovician orogeny in the area is proved by dated thrust-faulting and low-grade metamorphism in the Hudson Valley not far to the northwest.

The Spruce Pine pegmatites lie in a curiously similar position, close to the southeast edge of the certainly Precambrian core of the Blue Ridge welt in metamorphic rocks of less certain age. Stratigraphic (but not as yet tectonic) evidence for the existence of a pulse of the Taconian orogeny in the southern Appalachians has been accumulating rapidly for several years (King, 1950, pp. 660-661), and it is no longer necessary to explain the radioactive ages at Spruce Pine as Carboniferous as Holmes did in 1931 (p. 342). The orogenic pulse in the southern Appalachians was apparently Medial rather than Late Ordovician, however.

The other minerals dated in this range lie southeast or east of these well established groups. Little positive is known about the ages of the granites in the belt of the Piedmont that includes Marietta, S. C., and Statesville, N. C.; this is the belt next southeast of the Brevard zone. The Greenwich, Mass., allanite is supposed to be related to the Monson gneiss, and, as already noted, it lies on strike between the younger minerals around Middletown, Conn., and at Grafton, N. H. The north end of the belt of gneiss passing through Greenwich is compared by J. B. Hadley (personal communication, 18 July 1951) to the Bethlehem gneiss of New Hampshire. In places, however, the igneous rocks of this belt do include older magma series than the New Hampshire magma series, and possibly the Greenwich allanite comes from one of these. The Fitchburg, Mass., uraninite is supposed to be related to the Fitchburg granite, which cuts schists variously dated as Devonian or Carboniferous, but hardly older. On the other hand, the Blue berry Mountain pegmatite in Woburn, Mass., is supposed to be related to the Salem gabbrodiorite and Dedham granodi-

orite, members of one of the older igneous series of the Boston area, and these have been dated on geologic grounds as either early Paleozoic (LaForge, 1932, pp. 22, 27-28) or more probably Precambrian (Billings, 1929, p. 103; Dowse, 1950). Thus there is difficulty in squaring all three of these eastern Massachusetts absolute ages with the surrounding geology, and because for none of them was enough material available to check the degree of contamination with common lead by determining either the atomic weight or the isotopic constitution of the lead, they cannot be considered satisfactory.

PRECAMBRIAN DATES

Age determinations.—Several chemical analyses give dates of about 600 million years, as follows:

Class	Pb	Th	U	y	x	Age (million years)
Monazite from Mars Hill, Madison County, North Carolina (Marble, 1936):						
II	0.152	5.526	0.019	0.0274	0.003	600
	0.175	5.596	0.016	0.0312	0.003	680

An unpublished atomic weight determination by J. P. Marble indicates that very little if any non-radiogenic lead is present (Marble, personal communication, 9 September 1951).

(Lane, 1937, p. 58; on same material):						
IV		5.75	0.031	0.0287	0.005	620
	0.166	5.54	0.035	0.0298	0.006	650
Monazite from Deer Park mine (no. 5), Spruce Pine, Mitchell County, N. C. (Bliss, 1944):						
III	0.129	4.86	0.017	0.0265	0.003	580
Monazite from Topsham, Maine (Lane, 1937, p. 58):						
IV	0.168	5.89	0	0.0285	0	630

Others give dates of about 800 million years, as follows:

Allanite from northern Amherst County, Virginia (Marble, 1935, p. 351):						
III	0.056	1.245	0.080	0.0423	0.060	820
	0.060	1.254	0.083	0.0449	0.062	870
	0.062	1.251	0.087	0.0463	0.065	890
(Hecht and Kroupa, 1936, p. 101; on same material):						
IV	0.051	1.195	0[5]	0.0427	0	940[5]
Uranoan thorianite from Easton, Penna. (Wells and others, 1933, p. 50):						
III	4.96	33.87	31.50	0.0759	0.482	790
	4.72	33.75	31.05	0.0728	0.479	770

[5] The analytical method used in this analysis would not have detected 0.08 U. If the average of Marble's determinations of U is assumed, the age becomes 770 million years!

The data strongly suggests two periods of intrusion in the Precambrian, roughly 600 and 800 million years ago, but of course each of these periods may have included several pulses at considerable intervals.

Some helium determinations on magnetite from Precambrian rocks may be listed, though they scatter badly and can hardly be considered conclusive (Hurley and Goodman, 1943, pp. 310-311; Hurley, 1949, discusses these discrepancies and presents the results of an alternative method, which however gives much younger ages and is considered even less satisfactory):

Ringwood, N. J.	410	million	years
Ford mine, Stanhope, N. J.	550	"	"
Franklin Furnace, N. J.	640	"	"
Cranberry, N. C. (2 determinations)	830	"	"
	1260	"	"
Cumberland, R. I.	1500	"	"

Perhaps the New Jersey dates belong to the 600 million year range and the North Carolina dates to the 800 million year range, but little weight can be placed on them. There is no independent evidence to suggest that the cumberlandite at Cumberland, R. I., is so early in the Precambrian; indeed several observers have considered it Paleozoic (Emerson, 1917, pp. 183-184). In the absence of such other evidence, therefore, the figure of 1500 million years is here ignored.

Geologic setting.—The Spruce Pine and Topsham monazites come from pegmatite districts other minerals from which have yielded well attested younger dates. Although it is possible that there are two ages of pegmatites in these districts, no confirming evidence is known; J. M. Parker (personal communication, 29 October 1951) doubts it for Spruce Pine, and Shainin (1948, 1949) for Topsham. Shainin prefers the later Paleozoic age for Topsham on geologic evidence. The explanation of these anomalies is not forthcoming, unless common lead is present in large amount in the minerals.

Except for these and for Cumberland, R. I., the localities cited lie in the discontinuous belt of Precambrian anticlinoria along the western margin of the crystalline Appalachians—the "Archean protaxes" of Dana. Easton and the New Jersey localities lie in the southwest extension of the Hudson Highlands, the others in the Blue Ridge welt. Moreover, Spruce Pine lies at the southeast edge of this welt. The absence of Precambrian dates farther east hints perhaps that Precam-

brian rocks (except those of the latest Precambrian) are uncommon in the Piedmont, but age determinations are still too sparse there for this negative evidence to carry any weight.

TRIASSIC DATES

Nine helium determinations on magnetite in or associated with Triassic diabase intrusives or lava flows are listed by Hurley and Goodman (1943, p. 309). Four of these, based upon refined technique, give the following results:

Garrish Mountain, Cochester County, Nova Scotia .... (from magnetite deposit associated with flow)	158	million	years
Lakeville, Kings County, Nova Scotia ..... (from veinlets in a flow)	169	"	"
Cashtown, Adams County, Pennsylvania ..... (from magnetite deposit associated with intrusive)	174	"	"
"Virginia" (presumably west of Washington, D. C.) .. (from intrusive)	165	"	"

The other five, made on imperfectly purified material, give minimum ages only:

Preakness Mountain, Passaic County, New Jersey ..... (from second Watchung flow)	more than	153	million	years	
French Creek, Chester County, Pennsylvania ..... (from magnetite deposit associated with intrusive into Precambrian rocks)	"	"	140	"	"
Boyertown, Berks County, Pennsylvania ..... (from magnetite deposit associated with intrusive)	"	"	152	"	"
Dillsburg, York County, Pennsylvania ..... (from magnetite deposite associated with intrusive)	"	"	143	"	"
Goose Creek, Loudoun County, Virginia ..... (from basic pegmatite associated with intrusive)	"	"	140	"	"

In an earlier paper (1941, p. 554), Hurley and Goodman give a determination from Cornwall, Lebanon County, Pennsylvania, presumably a minimum, of 126 million years.

These results show a gratifying consistency, especially in contrast to the helium determinations on Precambrian magnetites. The approximate mean age of the Triassic igneous activity may therefore be taken as 170 million years or a shade less.

Colbert (1946, pp. 266-271) has recently reviewed the geologic evidence on the age of the Triassic rocks of eastern North America and concluded that their age is Late Triassic, probably fairly early in the epoch.

SUMMARY

Except for determinations on Triassic minerals, the available absolute age determinations in the Appalachian region

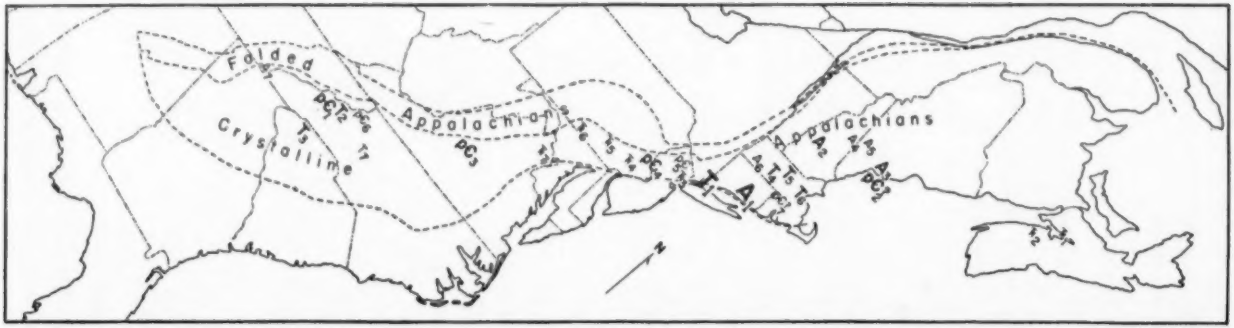

Fig. 1. Localities of Appalachian radioactive minerals on which absolute age determinations have been made. Large heavy letters indicate consistent groups of lead-ratio determinations including at least one lead isotope analysis; moderate letters indicate other lead-ratio determinations; small light letters indicate other types of determinations. A—later Paleozoic dates (Acadian, Appalachian?); T—earlier Paleozoic dates (Taconian); pC—Precambrian dates; Tr—Triassic dates.

Localities indicated are:

A_1, vicinity of Middletown, Conn.
A_2, Grafton, N. H.
A_3, Topsham, Me.
A_4, Conway, N. H.
A_5, Paris and vicinity, Me.
A_6, Chester, Mass.
A_7, Ducktown, Tenn.

T_1, Bedford, N. Y., and Branchville, Conn.
T_2, Spruce Pine, N. C.
T_3, Marietta, S. C.
T_4, Greenwich, Mass.
T_5, Fitchburg, Mass.
T_6, Woburn, Mass.
T_7, Statesville, N. C.

pC_1, Mars Hill, N. C.
pC_2, Topsham, Me.
pC_3, Amherst, Co., Va.
pC_4, Easton, Pa.
pC_5, Highlands of New Jersey
pC_6, Cranberry, N. C.
pC_7, Cumberland, R. I.

Tr_1, Garrish Mountain, N. S.
Tr_2, Lakeville, N. S.
Tr_3, Watchung Mtn., N. J.
Tr_4, Chester and Berks Cos., Pa.
Tr_5, Dillsburg, Pa.
Tr_6, Cashtown, Pa.
Tr_7, Goose Creek, Va.

fall generally into four groups according to age, suggesting major orogenies about 800, 600, 350, and 260 million years ago. (Each orogeny may have comprised several local pulses, differing somewhat in date from area to area, as indeed the later ones are known to do.) The two earlier orogenies were Precambrian, and evidence for them is confined to the belt of Precambrian uplifts at the west edge of the crystalline Appalachians. The next youngest is presumed to be Taconian (Ordovician); it affected the western part of the crystalline Appalachians in the Carolinas and New England and also perhaps part of eastern Massachusetts. The last well recorded by radioactivity is the Acadian, which affected eastern New England, especially a belt east of the Connecticut River, perhaps in Mississippian time. Surprisingly, the Appalachian orogeny, which deformed the Valley and Ridge province and the coal basins of Rhode Island and the Maritime Provinces, probably in Permian time, is unrepresented by published absolute age determinations, unless perhaps by helium determinations on ore minerals from Tennessee and western Massachusetts or by determinations on pegmatite minerals from southern Maine.

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A NEW QUANTITATIVE METHOD OF TOPOGRAPHIC ANALYSIS

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ABSTRACT. A new statistical method of topographic analysis has been developed and shows good promise of becoming useful in a wide variety of problems. The method is simple to apply and makes use of what is called the statistical slope orientation diagram. A number of topographic characteristics may be revealed and studied with the aid of this diagram. Some of these characteristics are not brought out by other methods of examination. The general application of the diagram to geologic studies seems well established but the method needs further investigation to determine its limitations. Instructions for preparing and interpreting the diagram are given together with some actual examples. A number of suggestions for further study are mentioned.

INTRODUCTION

A new method of quantitative slope study has been investigated and makes use of what is called the statistical slope orientation diagram. The method has been only very briefly presented (Chapman, 1951) but further study indicates that enough information has now been accumulated to warrant a detailed account of this statistical method of analysis. Hereafter, the statistical slope orientation diagram will be referred to as the SSO diagram.

In brief, the SSO diagram is a statistical representation of the orientation in space of the various slopes of a given area. It is analogous to a petrofabric diagram in which the orientation of mineral grains within a rock are shown statistically. The slope of the land is first measured at a great number of places on a topographic map, and these slope values are plotted as points on an equal area projection to give what is called a pole diagram. One hypothetical diagram is shown in figure 2. The concentration and distribution of points is more clearly demonstrated by contour lines. This contoured diagram is the completed SSO diagram, and an actual example is given in figure 4. From diagrams of this type it is possible to determine quantitatively the distribution of slopes with reference to direction and angle of inclination.

Before considering the preparation and use of the SSO diagram, the characteristics of this analytical method will be enumerated.

1. Direction of slope is shown to be as significant as angle of slope.
2. Frequency of a given slope orientation is proportional to the map area underlain by that slope.
3. Absolute and relative elevations are ignored.
4. Geographic position is ignored.
5. Slopes are studied on an areal basis and not a cross-sectional basis.
6. The method helps to obtain a true picture of slope relations. This picture is not over-influenced by features of small areal extent which may otherwise appear outstanding.
7. The method reveals characteristic differences in land surfaces, which are not made conspicuous by other methods.
8. Data may be selected so that certain features of topography may be emphasized or suppressed as the situation warrants.
9. As a result of certain characteristics listed above, a few SSO diagrams may be heteromorphic. That is, one diagram may represent two or more different types of topography. Most diagrams, however, will probably prove to represent a single topographic type and may be referred to as monomorphic.

METHOD OF PREPARING THE SSO DIAGRAM

Method of sampling.—The successful application of the SSO diagram depends upon an accurate determination of the relative map areas underlain by slopes for all angles of inclination and strike. Careful consideration, therefore, must be given to the problem of obtaining accurate, statistical samples. A simple method of sampling the slopes of a given map area is to run a series of uniformly spaced parallel traverses along which slope measurements are taken at regular intervals. This method may be referred to as the traverse method; the spacing of traverses and measurements along traverses will be determined by the topography and by the accuracy desired. A fine and uniformly textured topography requires fewer traverses than a coarse or non-uniform one. The spacing of measurements along traverse lines may also vary with texture and uniformity of topography. The direction in which trav-

erses are to be run is an important problem to consider, particularly if the topography exhibits a directive texture. A linear texture, such as is common in areas of folded sediments, will be most accurately sampled, by the traverse method, if traverses are run normal to the ridge and valley lines.

Let us assume that about 100 measurements are necessary in order to statistically sample a map area 5 x 5 inches square. Using the traverse method, we might run two traverses with measurements taken every tenth of an inch apart; we might run three equally spaced traverses with measurements taken every seventh of an inch apart; or we might choose five traverses with measurements spaced every quarter of an inch. For certain types of topography closely spaced measurements on fewer traverses may be better, whereas in most cases a larger number of traverses is to be desired. If we continue to decrease the traverse spacing and at the same time increase the distance between the regularly spaced measurements on traverse lines, we will reach a condition where spacing of measurements is equal in two directions at right angles. This gives a square grid-like arrangement of points and sampling by this pattern is called the rectangular grid method.

The writer believes the rectangular grid is only slightly more time-consuming to employ but it offers one outstanding advantage. Less attention need be given to orienting the rectangular grid with respect to a directive texture of the topography. This point is particularly important in cases where the directive texture has a different orientation in different parts of the area. One example of this may be found in an area where plunging folds are expressed topographically and alternating limbs have distinctly different trends.

To carry this a step further, a hexagonal grid could be used. In such a case the points of measurement are equally spaced along three sets of lines which intersect at 60° angles. Theoretically this method eliminates more completely the problem of directive texture because net points are more radially spaced. It is questionable, however, as to whether the increased accuracy obtained by its use is significant in view of certain inherent errors in the whole method of analysis. The hexagonal grid method is somewhat more difficult to employ than either of the other two methods already described.

If the traverse method is used, parallel traverse lines may

be drawn across the map in the most advantageous direction. The uniformly spaced points, at which slope measurements are to be taken, are then indicated by small tick marks on the traverse lines. If the rectangular grid method is employed, the intersections of the two sets of perpendicular lines in the grid are indicated on the map by points or small crosses.

Having chosen the most satisfactory method of sampling, the next step is to proceed systematically across the map measuring the strike and dip values for the slopes at each point indicated. The strike of a slope is readily measured to within an error of a few degrees by determining the direction of the tangent to the contour line at the point of measurement. If the point does not fall on a contour line, an average or interpolated value is determined from the two adjacent contours. This strike value is recorded on the map just above the point as in the following examples: N10E, N55W, N, N90E, etc. The angle of slope is next determined by using a special slope scale which will be explained shortly. The slope angle and general slope direction are recorded on the map just below the point as in the following examples: 20°W, 30°E, 5°S, etc. The representation is, thus, analogous to strike and dip of bedding planes.

The slope scale.—The slope scale is a device to determine slope angle from contour spacing. The scale, as shown in figure 1, is prepared from a small piece of transparent celluloid. A number of fine lines of various lengths are scratched with a sharp point on the underside of the celluloid. The lengths of the lines equal the spacing of topographic contours for different angles of slope and may be determined for each slope angle desired by the following formula:

$$S = VR \text{ cotan } A$$

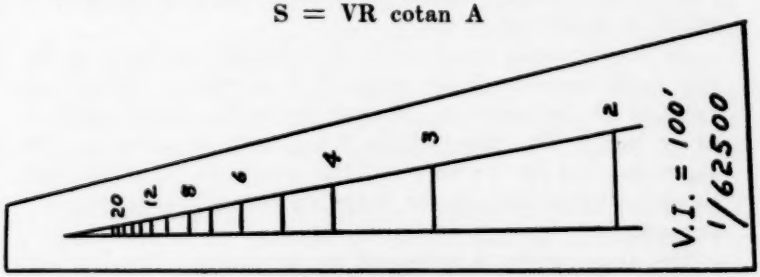

Fig. 1. The slope scale.

where S equals contour spacing in inches, V equals the vertical interval (in inches) between the contours to be measured, R is the representative fraction of the map and A equals the angle of slope.

It is a difficult matter to lay off accurately a series of lines whose lengths are equal to the various S values determined from the above formula; therefore, the following scheme is suggested. On a sheet of paper carefully construct two straight lines to form an angle of 10°. These two lines are represented in figure 1 by the sides of the slope scale. Each S value is next multiplied by the tangent of 10° and the lengths obtained are carefully laid off along one side of the 10° angle by making fine tick marks. All measurements are made from the vertex of the angle. A fine line is now drawn from each tick mark, in a direction perpendicular to the side of the angle, until it intersects the other side of the 10° angle. If construction has been accurate, the sides of the 10° angle will cut off intercepts, on the short parallel lines, equal to the S values computed above. The lengths of these lines may be checked against the theoretical values.

The piece of thin celluloid is now placed over the 10° angle and all lines are carefully scratched on the celluloid with a fine point. Care should be taken in this process to avoid the effects of parallax and refraction. After each parallel segment has been labelled with its proper degree value and a record made of the map scale and contour interval used, the slope scale is ready for use. In practice it may be found convenient to prepare two slope scales, one for measuring spacings of adjacent contours in the areas of low relief and the other for spacings of every fifth contour where relief is high. Other combinations of scales may be found more suited to particular areas.

To use the scale merely slide it along the contours on the map until the sides of the wedge just fit between the chosen contours at the point on the map where the measurement is to be taken. The slope of the topography is given by the degree reading on the wedge at this point. If the point falls between labelled lines on the wedge then an interpolation may be made.

The slope strike is measured by determining the bearing of a line tangent to the contour line at the point of measure-

ment. This angle may be found with a protractor. The strike and dip of the slope should be recorded on the map at the point of measurement. This may be done in ink for a permanent record. This method of recording aids checking data later, and it enables one to prepare selective diagrams if desired.

A large reading glass with about a 2X magnification and mounted on a long arm with pedestal greatly facilitates slope measurement.

Plotting.—In order to get a visual picture of all the slope data, it is necessary to imagine that each set of slope measurements represents a plane surface tangent to the topographic surface at the point where the set of measurements was taken. If all these tangent planes, for the numerous points measured, are visualized as being grouped at the center of a large sphere and if a straight line is drawn perpendicular to each plane from the center of the sphere, then each line extended will cut the sphere at two diametrical points. The sphere will be covered with a great number of points, each diametrical pair of points having its position determined by the orientation of the corresponding tangent plane in space. With this understanding, two points may be used to represent a tangent plane or surface of slope. A horizontal slope would be represented by a point at the top and bottom of the sphere; a vertical slope by two points diametrically located on the equator; and inclined slopes by two points, one above and the other below the equator. It is now clear that all points on the upper hemisphere will have counterparts on the lower hemisphere. Consequently only the upper hemisphere may be considered; and it may be said, therefore, that the orientation in space of each tangent plane or slope surface may be represented by a single point on the sphere. This representation is called spherical projection and each plane is represented by a point which is called the pole of the plane. There is one exception to the above statement. Vertical slopes or tangent planes will be represented by two poles on the equator diametrically opposed.

Spherical projection of data requires a three-dimensional model but it is possible to obtain spatial relations by a simplified two-dimensional projection somewhat similar to those used in crystal projection and map-making. Each slope

surface (tangent plane), as before, is represented by a point (pole) in the projection. When the poles of the numerous slope surfaces are plotted a pole diagram results. Figure 2 represents a hypothetical pole diagram. It is a sort of top view (in two dimensions) of the upper hemisphere. Pole A represents a horizontal plane. Poles B and B^1 represent a vertical plane striking N-S. Poles C and C^1 represent a vertical plane striking E-W. Poles D and D^1 represent a vertical plane striking NE-SW. Pole E represents a plane striking N-S and dipping 50° E. Pole F represents a plane striking N 30 W and dipping 70° E.

The type of projection prepared for the present study is known by such names as Lambert equal area, equi-areal equatorial, and surface-true azimuthal projection. This type of projection is used in order to compare unit areas which remain unchanged in different parts of the projection diagram. The plotting of data is greatly facilitated by use of a prepared projection net which is a type of scale for plotting data directly on a sheet of paper. These nets are employed in plotting data for petrofabric analysis and are commonly referred to as Schmidt nets. They can be supplied commercially in the standard size of 10 centimeter radius.

Slope angles seldom exceed 30° and, therefore, only the

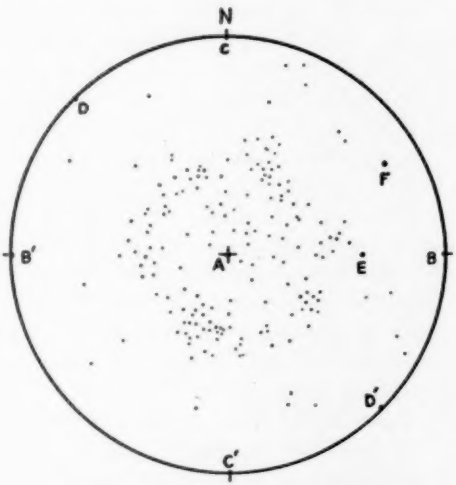

Fig. 2. Hypothetical pole diagram.

central part of the equal-area net will be used. This means that, with the standard 10 centimeter net, nearly all points on the diagram will fall within the 30° circle; and poles will be crowded within such a limited area as to be confusing. It is advisable, therefore, to use only the central portion of the projection net and to enlarge it several times thereby simplifying the plotting problem. The writer has found that by enlarging the central portion of the net to a 10 centimeter radius, dip values up to 30° may be plotted very satisfactorily; and the net can be readily mounted on the same metal turntable[1] used for stereographic projection and petrofabric analysis. Net enlargements are not difficult to construct particularly if prepared in a skeletal form as shown in figure 3. This may be called a 30° net because the horizontal line is divided into 1° segments up to 30°. The 30° circle is divided into 2° segments and numbered from 0° to 90° in either direction. The index or North direction coincides with 0°. These are the only elements of the net which are es-

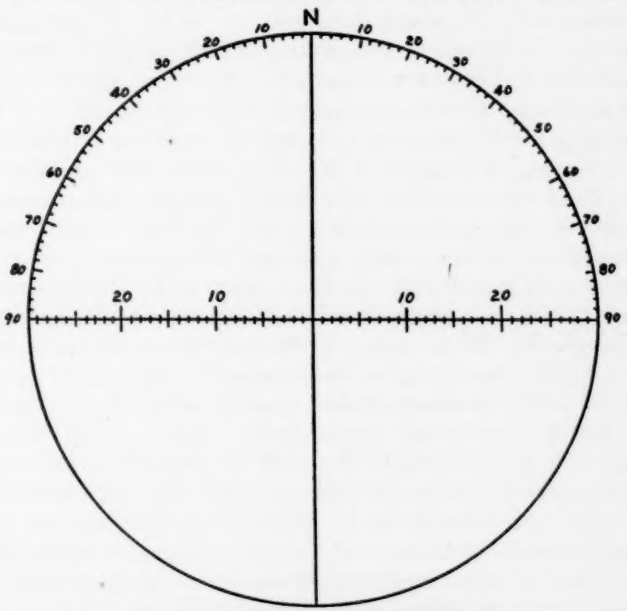

Fig. 3. Skeletal equal-area projection net.

[1] This turntable is supplied by E. Leitz Co., New York.

sential for plotting poles. A special net may be prepared for plotting values up to say 45° should the need arise. The 10 centimeter circle with 2° subdivisions is used but the horizontal line is subdivided into segments up to 45° instead of 30°. In the same manner if the horizontal line is subdivided into 1° segments up to 10° or 15°, a skeleton net is found which is ideal for topography where slopes are unusually low. A 15° net gives about twice the enlargement of the 30° net and a 10° net enlarges about three times that of the 30° net.

The general formula for determining the length of subdivisions along the horizontal line of the net is:

$$r = \sqrt{2}\, R\, \frac{a}{2}$$

where r is the distance along the horizontal line from the center of the projection, R is the radius of the sphere or 90° circle, and a is the angle to be represented. It might be mentioned that the spacing along the horizontal line is essentially uniform for angles up to about 20°, and the same is true for all practical purposes for angles up to about 30° where the r value for 30° does not exceed about 10 centimeters.

To plot values with the turntable device, a sheet of transparent paper is placed over the net and attached to the outer movable ring. The circle of the net is traced through onto the paper and on this circle the North direction index mark is located and labelled. If desired other cardinal points may be shown. The movable ring with paper is rotated until the North index comes opposite the correct angle on the divided circle, which represents the strike angle of the slope surface to be plotted. Strike values to the east of north are located by rotating the paper counterclockwise and values to the west of north by rotating the paper clockwise. Having turned the North index to the proper strike angle, the dip value is located on the horizontal line, and a point is made on the paper directly above the correct dip value. Easterly dip values are measured to the right of the projection center and westerly values to the left of the center. Successive poles may be plotted by rotating the North index on the paper to the proper strike angle and making a point on the paper at the proper dip value.

Instead of using the metal turntable, a less expensive but

equally satisfactory device may be prepared by cementing the net to stiff cardboard or plywood and covering it with a thin sheet of celluloid for protection. A piece of transparent paper is fastened to this device by a pin which is inserted at the exact center of the net. The paper may thus be rotated about the pin to any desired position. If a small square of scotch tape is fastened to the underside of the transparent paper, enlargement of the pinhole will be avoided and yet the paper is rendered no less serviceable. The method of operating this device is precisely like that for the metal turntable.

Contouring.—The concentration or density distribution of poles, as revealed by the pole diagram, is not always easy to determine. Density relations may be shown more quantitatively by use of a contour diagram. This diagram is referred to as the *statistical slope orientation diagram* or briefly the SSO diagram. To prepare the contour diagram, the pole diagram is placed on a sheet of square grid paper. The size of each grid square should be one centimeter on a side. The center of the pole diagram should coincide with one of the line intersections of the square grid. A clean sheet of transparent paper is next placed on top of the pole diagram so the grid lines and the points on the pole diagram are visible through the upper sheet.

A small counter is prepared by cutting a circular hole (one centimeter radius) in a small square of cardboard. A better counter may be prepared by scratching a circle (one centimeter radius) on one side of a small square of transparent celluloid. In either case, the circle or circular hole has an area equal to one percent of the area of the pole diagram. It is now possible to sample the concentration of poles at different points in the diagram by placing the counter at the desired position and counting the number of poles which fall within the small circle. This number divided by the total number of poles in the entire diagram gives the percentage concentration at the point measured. Density data may be obtained for the entire diagram if samples are taken systematically by placing the counter at each grid intersection, counting the number of poles within the small circle, and recording their number at the grid intersection (at the center of the small circle). The diagram may be systematically

covered again, this time placing the center of the counter circle at the center of each little grid square and recording there the number of poles falling within the small circle. These additional values render the final contour diagram somewhat more accurate and are considered well worthwhile.

Contour lines or lines of equal pole density may now be shown similar to the way topographic contours are drawn from points of known elevation. The one per cent contour should pass through numbers on the sheet which represent one per cent of the total number of poles. The two per cent contour should pass through density numbers which are two per cent of the total number of poles, etc. For further details the reader is referred to Knopf and Ingerson (1938, pp. 245-251) or Fairbairn (1949, pp. 275-290). It is worthwhile noting that, since slope angles as high as 80° or 90° will almost never be encountered, one will not need to work in the outer limit of the projection as is true in petrofabric studies. Consequently the special periferal counter used for petrofabric diagrams has no application here. Actually we are working with only part of the complete hemisphere, and the 30° circle in our diagram is merely an arbitrary reference for purposes of standardization. The choice of a one per cent counter (one per cent of the 30° circle) is merely another means of standardizing the diagrams. Actually if any poles fall outside our 30° circle, a slight inconsistency is introduced by applying the one per cent counter to this area. If the number falling outside is small, then no serious error is encountered. If this number is great, however, it may be desirable to select a larger reference circle (i.e., 45°).

The completed SSO diagram should be properly indexed with the north direction; and data as to geographical location, number of poles represented and value of contours should be given. See figure 4.

ELEMENTS OF THE SSO DIAGRAM

Isogonic line—a radial line in the projection all points of which represent the poles of slopes with the same slope direction. Each radius of the projection is a different isogonic line.

Isogonic zone—that part of the projection enclosed by two closely spaced isogonic lines. All poles lying within a narrow

pie-shaped segment of the diagram lie within the same isogonic zone.

Isoclinal line—a circle all points of which represent the poles of equally inclined slope surfaces. Any circle concentric with the projection is an isoclinal line.

Isoclinal zone—that part of the projection enclosed by two closely spaced isoclinal lines. All poles lying within the relatively narrow band formed by two isoclinal lines lie within the same isoclinal zone.

Maximum—a relatively small area of the diagram within which the concentration of surface poles is unusually high. Most maxima are roughly circular.

Partial girdle or incomplete girdle[2]—a belt-like or band-like concentration of poles extending across the diagram in the general direction of a great circle. Each belt or band should pass through the center of the diagram and coincide with two diametrically opposed isoclinal zones. A partial girdle is formed by a great number of poles which represent slope surfaces with the same strike but widely ranging angles of inclination.

Small circle girdle—an uninterrupted circular belt or roughly circular band with a high concentration of poles. Many small circle girdles are concentric with the projection; and, therefore, coincide with isoclinal zones. In some diagrams small circle girdles may be off centered by a number of degrees.

Partial small circle girdle or incomplete small circle girdle—a broken or highly interrupted small circle girdle. It may be coincident with an isoclinal zone or it may be off-centered.

SYMMETRY OF THE SSO DIAGRAM

General statement.—The symmetry pattern of the SSO diagram is dependent largely upon the drainage pattern; and this, in turn, is controlled by the type of erosive agent, the climate, the dominant rock structure, etc. Various combinations of these factors give rise to various types of symmetry,

[2] Objection may be raised to these terms because of the small portion of a great circle along which these bands can extend. A complete girdle around the sphere is, of course, impossible. The attempt is made here, however, to use the same nomenclature as that employed in petrofabrics.

but for convenience it is suggested that the following simple classification be employed. The symmetry types given below are more or less gradational; one type may pass gradually into that of any other. In many cases, therefore, individuals may differ as to which type of symmetry a particular diagram possesses. If this gradational character is kept in mind, however, the simplified scheme suggested here will serve a useful purpose.

Axial symmetry.—Axial symmetry is characteristic of those diagrams in which most any diameter divides the figure into essentially symmetrical halves. In other words, there are a great number of planes of symmetry intersecting at the center of the projection.

Orthorhombic symmetry.—If two mutually perpendicular symmetry planes intersect at the center of the figure, then orthorhombic symmetry exists.

Monoclinic symmetry.—This is more common than either of the above. It is characterized by a single plane of symmetry.

Triclinic symmetry.—If no planes of symmetry exist, the diagram shows triclinic symmetry. This is perhaps the most common symmetry type.

ELEMENTS OF TOPOGRAPHY REVEALED BY THE SSO DIAGRAM

General statement.—Perhaps the greatest value of the SSO diagram is that it permits a study of certain elements of topography not clearly revealed or not revealed at all by other methods of analysis. A number of these elements together with others which may be investigated with the aid of pole diagrams and SSO diagrams will now be considered.

Slope direction and slope angle.—Slope direction or the angle between the north line and the direction in which a surface inclines is determined from the diagram by measuring the direction of the isogonic line through the pole of the specified slope. The direction of slope is at right angles to the strike of the slope. The slope angle or angle of inclination of a slope surface may be determined from the diagram by measuring the angular distance between the center of the diagram and the desired slope pole. This may be done by using the subdivided equator line as a ruler.

A useful curve may be prepared from the slope values shown in the pole diagram. First, the diagram is divided into a number of pie-shaped segments by a series of uniformly spaced isogonic lines. Next, the average slope angle for all points falling within each segment is computed. This average is plotted on the central isogonic line in each segment, at the proper distance from the center of the diagram. The newly plotted points are now joined by straight lines to give what may be called the *average slope angle curve.* See figure 5. It should be noted that the values are not weighted averages for the diagram as a whole but merely indicate the average slope of land within a specified isogonic zone. Further use of this slope angle curve will be made later.

The pole diagram offers a convenient means of computing the percentage map area underlain by slopes of any specific angle of inclination or strike. For example, to determine the percentage of the map area underlain by slopes whose angles of inclination lie between 5° and 10°, merely count the poles between the 5° and 10° isoclinal lines and divide the number by the total number of poles in the diagram. Similarly the number of poles between any two isogonic lines divided by the total number of poles in the diagram gives the percentage of the map area underlain by slopes whose directions lie within the range of the isogonic lines chosen. It may be desired to determine the percentage of map area underlain by slopes falling within specified areas in the diagram. In this case the poles are counted within the area selected and the number is divided by the total number of poles in the diagram to get the percentage desired.

Dispersion of slope angles.—The amount of difference between the extreme slope angles of a diagram is the dispersion of slope angles. This dispersion may also be used to express the angular difference between slope angles within specified directions or parts of the diagram. These values could be obtained readily from the pole diagram but such a practice is not considered advisable. On most diagrams a few scattered poles lie outside the lowest contour line and, therefore, represent statistically insignificant values. It seems advisable, in view of this, to obtain values from the contour diagram.

Dispersion of slope direction.—The angle between the two

isogonic lines which bound the more or less continuous distribution of poles is the dispersion of slope direction. In most areas slope directions box the compass so that the dispersion is 360°. When applied to the diagram as a whole, therefore, this dispersion value is of little worth. It is useful, however, when applied to specific parts of the SSO diagram such as the portion bounded by two isoclinal lines. In this case the dispersion is much more limited. To determine the dispersion values, the contour diagram and not the pole diagram should be used.

Frequency of slope angles.—The frequency with which each angle of slope is encountered, regardless of the slope direction, is the slope angle frequency. It is computed by noting the changes in pole density within successive isoclinal zones. This frequency may be crudely estimated by observing the distribution of contour lines. A more accurate relation may be obtained from the pole diagram by counting the number of poles falling within successive isoclinal zones. A frequency curve may be prepared by counting the number of poles between the 0° and 2° isoclinal lines and then counting the number between the 2° and 4° isoclinal lines and so on until values for each 2° isoclinal zone are obtained. By plotting slope angles as abscissae and pole numbers (frequencies) as ordinates, a *slope angle frequency curve* is obtained. This curve shows the relative map areas underlain by different angles of slope but it completely ignores slope direction. Nevertheless, the slope angle frequency curve is a diagnostic character of each type of topography. No examples of these frequency curves are given in this paper.

Frequency of slope direction.—The frequency with which each direction of slope is encountered, regardless of the slope angle, is the slope direction frequency. It is computed by noting the changes in pole density within successive isogonic zones. It is not difficult to estimate this frequency by observing the SSO diagram but a curve may be prepared from the pole diagram by counting the number of poles within each two-degree isogonic zone. Zones of greater angular value may be preferred. These frequency numbers are plotted as ordinates against the various direction angles ranging from 0° to 360° to complete the *slope direction frequency curve*. Another way

is to plot the values radially (fig. 5) in the same manner as the average slope angle curve. Such a curve has certain quantitative value but it ignores slope angles. It shows at a glance the relative map area underlain by slopes within each isogonic zone.

Symmetry of slope angles.—The relative values of average slope angles on opposite sides of the diagram center determines the symmetry of slope angles. If the average slope angle within one isogonic zone is essentially the same as that for the opposite zone, then symmetry of slope angles exists for those two zones. If the values are not the same, then there is asymmetry of slope angles in those zones. Within the entire SSO diagram, symmetry may exist between certain isogonic zones and not between others. These symmetry relations may be quickly determined from the average slope angle curve (fig. 5).

Symmetry of slope direction.—The relative map area represented by slopes within one isogonic zone as compared with that of the opposite zone determines the symmetry of slope direction. If the number of poles within diametrically opposite isogonic zones is approximately the same, then slope direction symmetry exists for those two zones because map area is proportional to the number of poles represented. In such a case there is as much map area inclining in one of the directions as in the other. Within a given SSO diagram, symmetry of slope direction may exist between certain isogonic zones and not others. Slope direction symmetry may be rapidly evaluated from the slope direction frequency curve (fig. 5).

Mean slope.—Mean slope is used to indicate the weighted geometric mean of all pole values in the diagram. Both direction of slope and angle of slope must be considered. A weighted value is obtained by using the total number of poles as a basis, and the resulting value, therefore, is a function of map area. The products of the corresponding values on the average slope angle curve and the slope direction frequency curve (fig. 5) represent the radial vectors for the *mean slope curve* (weighted average). See figure 5. The mean slope value may be quickly located by cutting a piece of cardboard to the same shape as the mean slope curve and then locating the center of gravity of this cardboard by

balancing it on the point of a needle. The point of balance is the location of the mean slope, whose direction and slope angle may now be readily determined.

Modal slope.—The mode represents the most frequently occurring slope value. In the SSO diagram it is represented by the highest possible contour line. Usually it is centrally located within the innermost contour line shown. It is defined according to direction and angle of inclination.

EXAMPLES OF SSO DIAGRAMS

The first example of an SSO diagram to be considered is one prepared from the topographic map of the New Bloomfield quadrangle, Pennsylvania. This diagram is shown in figure 4 and is based on 218 slope measurements taken from all but the northern and extreme southwestern parts of the quadrangle. The map used was the edition of 1907, the area being surveyed in 1904 on a scale of 1:62500 and a contour interval of 20 feet. Even though the number of poles is small, it appears to be sufficient to represent the essential topographic features.

The diagram shows a conspicuous and well-centered maximum reaching a concentration of about 22 per cent (mode) at the center. A conspicuous partial girdle extends in a direction roughly NNW-SSE and lies within the 22° isoclinal line. The symmetry is triclinic but only slightly off from monoclinic. The dispersion of slope angles is about 18° at one end of the partial girdle and 22° at the other end. Dispersion of slope direction remains constant (360°) for slopes between 0° and 10° but it decreases rapidly as the slope angle increases beyond 10°. The slope angle frequency decreases more or less continuously as slope angle increases, indicating an absence of sharp breaks in slope. The two highest frequencies for slope direction are roughly S 25 E and N 10 W (fig. 5). The implications of these two values will be considered later. The symmetry of slope angles is moderately good around the diagram. The symmetry of slope direction is only moderately good; the two highest frequencies for slope direction are not diametrically opposed but differ by about 15° from this position. In addition, westerly slopes are slightly more extensive than easterly slopes. The maximum

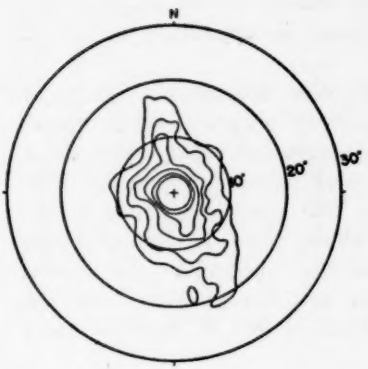

Fig. 4

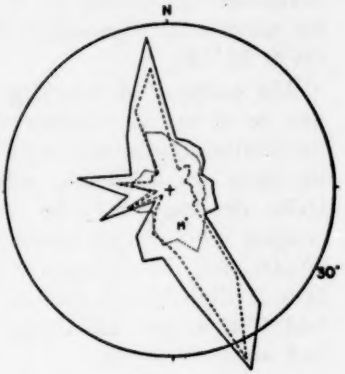

Fig. 5

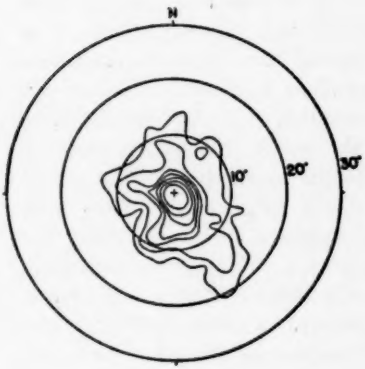

Fig. 6

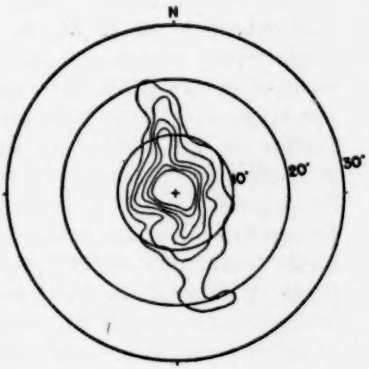

Fig. 7

Fig. 4. SSO diagram, New Bloomfield, Pennsylvania Quadrangle. Representing north and south limbs of fold; 218 poles; contours 2-4-6-8-10-12% per 1% area of 30° circle.

Fig. 5. Slope curves, New Bloomfield, Pennsylvania Quadrangle. Representing north and south limbs of fold. Dotted line, average slope angle curve; solid line, slope direction frequency curve; dashed line, mean slope curve; M, mean.

Fig. 6. SSO diagram, New Bloomfield, Pennsylvania Quadrangle. Representing north limb of fold: 126 poles; contours 2-4-6-8-10-12% per 1% area of 30° circle.

Fig. 7. SSO diagram, New Bloomfield, Pennsylvania Quadrangle. Representing south limb of fold, 96 poles; contours 2-4-6-8-10-12% per 1% of 30° circle.

concentration (mode) is at the center of the diagram but the mean slope (geometrical mean) is approximately 6° to the S 26° E.

The purpose of selecting the New Bloomfield quadrangle was to determine whether the SSO diagram would reflect the relationship between rock structure and topography. From the map, a conspicuous ridge and valley topography, with trellis drainage, may be seen. The general pattern is developed on a large syncline which plunges slightly north of east, and the direction of trend in topography and stream lines is slightly different on the two diverging limbs of the fold. The north limb trends about N 60° E and the south limb about N 83° E.

Close study of the SSO diagram shows that the partial girdle is not part of a great circle because the northern half extends in a direction about N 10 W whereas the southern half extends S 25 E. If the girdle were the expression of a single valley system one half should have been in diametric position to the other. The suggestion is, therefore, that two partial girdles are present, one trending N 10 W but extending farther to the north than to the south, the other trending N 25 W but extending farther to the south than to the north. Such girdles would be formed if the ridges and valleys were asymmetric in cross-section. It appears, therefore, that the girdle trending N 10° W represents a valley system trending N 80° E in which north slopes are steeper than south slopes. The girdle trending N 25° W represents a second valley system trending N 65° E in which south slopes are steeper than north slopes. Thus a valley system is represented at right angles to each of the two directions of greatest slope direction frequency and these correspond to the valleys on the north and south limbs of the syncline. The angular correlation between diagram and map is very close.

To test this relation further, an SSO diagram was prepared from the 126 slope measurements taken on the north limb of the fold, and a separate diagram was prepared from the 96 measurements obtained on the south limb. Figure 6 shows the SSO diagram for the north limb, and the partial girdle is seen to trend about N 34° W with steeper slopes to the south. Figure 7 is the diagram for the south limb and the partial girdle trends N 12° W. Except for a very few

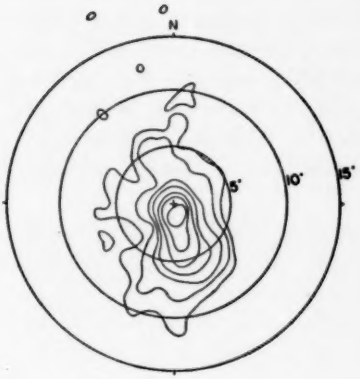

Fig. 8

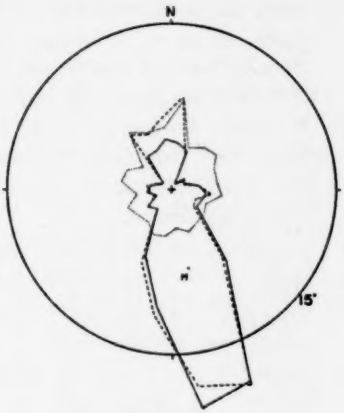

Fig. 9

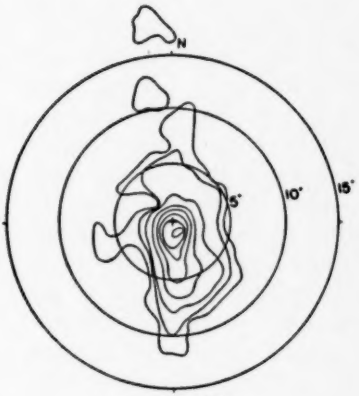

Fig. 10

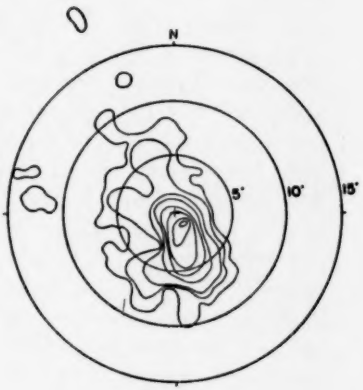

Fig. 11

Fig. 8. SSO diagram, Booneville, Arkansas Quadrangle. Representing odd and even-numbered traverses; 321 poles; contours 1-2-4-8-12-16-20% per 1% area of 15° circle.

Fig. 9. Slope curves, Booneville, Arkansas Quadrangle. Representing odd and even-numbered traverses. Dotted line, average slope angle curve; solid line, slope direction frequency curve; dashed line, mean slope curve; M, mean.

Fig. 10. SSO diagram, Booneville, Arkansas Quadrangle. Representing odd-numbered traverses; 160 poles; contours 1-2-4-8-12-16-20-24% per 1% area of 15° circle.

Fig. 11. SSO diagram, Booneville, Arkansas Quadrangle. Representing even-numbered traverses; 161 poles; contours 1-2-4-8-12-16-20-24% per 1% of 15° circle.

steep slopes the northward slopes are steeper than the southward slopes. The asymmetry of slope angles is much more pronounced on the north limb than on the south limb and the SSO diagram for the south limb is not far from orthorhombric symmetry. It is interesting to note that in each of these two diagrams the symmetry is essentially monoclinic, but when the data for both limbs are combined, triclinic symmetry results.

Another SSO diagram was prepared from the topographic map of the Booneville quadrangle, Arkansas. The edition used was that of 1939; the area was surveyed in 1933-34 on a scale of 1:62500 and a contour interval of 20 feet. Data were accumulated from about 50 square miles in the northern half of the quadrangle, in the area north of State Highway No. 10 and south of Backbone Mountain. This section shows several more or less continuous hogback ridges with considerably steeper slopes facing north. In preparing the SSO diagram shown in figure 8, 321 sets of slope measurements were obtained. Here the traverse method was used, and ten traverses were run (1¼ inches apart) in a north-south direction, about perpendicular to the trend of the ridges. Measurements were taken every eighth of an inch along the traverse lines. The diagram shows a strong maximum off-centered about 1° to the southeast, where it reaches a concentration of 23 per cent (mode). A poorly developed partial girdle trends roughly N 15° W and is at right angles to the trend of the hogback ridges shown on the map. The symmetry is triclinic but not far from monoclinic. Slope angle dispersion is much greater at the northern end of the partial girdle. Dispersion of slope direction is somewhat irregular where slope angles exceed about 5°. Slope angle frequency decreases fairly continuously as slope angle increases, except in the regions of steepest dips. The highest slope direction frequency is S 13° E; the second highest, but several times less, is about N 10° W (fig. 9). The symmetry of slope angles is good in all except the N-S to NNW-SSE directions where northerly slopes average about 2½° steeper than the opposing southerly slopes. The symmetry of slope direction is moderately good in all directions except N-S to NNW-SSE, where southerly slopes are three and a half times as numerous as northerly ones. The maximum concentration (mode) is about 1° to

the S 12 E, and the mean slope (geometrical mean) is approximately 7° to the S 12 E.

When the topographic map of this area is examined, the most conspicuous features to the eye are the steep northerly slopes of the hogback ridges. A very small percentage of the map area is underlain by these steep slopes, however; and in the SSO diagram they should play a rather minor role. The poles of the steep slopes are widely scattered over the NNW segment of the diagram and five separated patches with only one per cent concentration appear in this part of the diagram. The exact position of any of these low concentrations is not significant because the clustering of about three poles within the one per cent counter is all that is required for such a concentration to appear. The general location of these low concentrations is significant, however, because it is controlled by the average trend of the hogback ridges. In this respect the map and SSO diagram are in close agreement. The ratio of 321 poles to an area of 50 square miles, as used in this SSO diagram, might be considered higher than the minimum required for an adequate sample. The reason for taking such closely spaced measurements on the traverses was to insure a sufficient number of points on the narrow belts of steep northerly slopes. In this manner the steep northerly slopes were sampled at 23 different places which means that these steep slopes constitute only about 7.5 per cent of the map area. It appears then that a large number of poles must be obtained in order to sample the steep slopes adequately.

It is instructive to see what happens when we prepare two separate diagrams of the same area, one based upon the data obtained from the odd-numbered traverses and the other from the even-numbered traverses. The results are shown in figures 10 and 11. In most respects these diagrams closely resemble each other but in certain ways they are quite different. It should be noted that each of the five scattered one per cent areas in figure 8 are represented, in a complimentary manner, by similar areas in figures 10 and 11. Two new concentrations appear in one of the diagrams. The position and size of these low concentrations tend to exaggerate the differences between these two SSO diagrams. If we ignore these high angle slopes the two diagrams are very nearly alike. As they stand by themselves, neither represents the area statistically; together, however, they are believed to be adequately representative.

ERRORS INVOLVED

Some of the more important errors likely to be involved in this statistical method include map errors, errors in measurement, errors in contouring and errors in sampling. Map errors may be held to a minimum by working with the best maps available. If care is not taken, considerable error may be involved in measuring the direction of slope strike. This error is likely to be more prevalent in relatively flat country than in areas of greater relief. Errors arising from poor sampling of topography with a directive texture have already been implied. This is most pronounced, perhaps, when the traverse method is used. It has been suggested that, in general, a better sample will be obtained by using the rectangular or hexagonal grid method. Another and very serious error may be involved if the number of pole values is not in excess of the minimum required for a representative sample. No simple rule is known to the writer which will determine the number of points necessary. Here more specific information is needed. The minimum number will depend upon such variables as topographic texture, homogeneity, slope dispersion, etc.

CONCLUSIONS

Investigations so far have shown that the SSO diagram is a quantitative method for expressing numerous slope relations in a single condensed picture. Certain features, readily detected by observing the topographic map, may also be detected in the SSO diagram. Other features, not clearly expressed by a contour map, may be rendered conspicuous by the SSO diagram. This alone is sufficient reason to indicate that the SSO diagram may be helpful in making certain topographic studies. Much work needs to be done, and a few suggestions will be made as to how the method might be applied to future studies.

1. The method might be used to show a quantitative relationship between joint pattern and topography.
2. It might be used to indicate the relative control of bedrock and glacial action upon topography.
3. In areas where large scale maps are available, the method might be used to differentiate till sheets of different ages.

4. The method might serve to indicate the quantitative influence of rock structure upon topography. A close relation might be found between dip of bedding and average slope values. The matter of relative differences in slope on different fold limbs needs further investigation.
5. A series of diagrams representing various stages in the erosion cycle might bring out some significant relations. It might be possible to establish boundaries between stages or substages in the cycle by using percentages of map area falling within given slope values.
6. It seems possible that the method might serve as a means of differentiating physiographic units.
7. Employed on a regional scale, the method might reveal certain striking relations and trends.
8. The method will probably be found most effective when individual SSO diagrams represent relatively small areas. By restricting the areas in this manner, each diagram will represent a topographic unit more accurately. Larger areas on the other hand may give composite diagrams (i.e., Bloomfield quadrangle) and appear too generalized.
9. The method should prove helpful where it is necessary to know the relative areas of land whose slope angles fall within certain limits. Such data are useful in problems of geography, agriculture, flood control and water run-off.

It is not expected that the method will solve all problems nor perhaps any problems in every area. It must be applied with discretion. The method needs criticism; it needs further test and study and it needs more refinement. The concept upon which the application of these diagrams depends is simply that the slope pattern for a given area is the result of a specific combination of factors such as: climate, nature of underlying rock material, age of land surface, geographical location, and geological processes involved. The value of this method, therefore, will be determined by the success with which we are able to relate the numerous slope values to the various processes and conditions which produced them. This is a problem for the geomorphologist.

The writer wishes to express thanks to Dr. Jack E. Harrison for many helpful suggestions and for careful reading of the manuscript.

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INFRARED SPECTRA OF SOME ROCK-FORMING MINERALS

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ABSTRACT. Infrared absorption spectra of some common carbonates, sulphates, phosphates, and silicates are described. The anion dominates the absorption characteristics. Absorption bands are shifted to longer wave lengths as the cations in mineral groups become heavier, and as aluminum is substituted for silicon in the silicates.

The absorption of radiation in infrared frequencies indicates notable covalent bonding in the minerals described; hence they are not ideal ionic minerals.

INTRODUCTION

INCREASING interest in the mineralogical applications of infrared absorption spectra has been shown recently by mineralogists and sedimentary petrologists. Since the extension of the technique to powdered minerals (Keller, 1948), several laboratories have conducted studies of the infrared absorption characteristics of the clay minerals and other minerals and rocks. Mineralogists are still in the preliminary stages of compiling a library of reference curves that are prerequisite for the correlation of spectral characteristics and molecular or lattice structures, and for the identification of unknown minerals and mixtures. In this investigation, the absorption spectra in the 2-15 microns wavelength range of several carbonate, sulphate, phosphate, and silicate minerals have been determined, to place on record the infrared spectral characteristics of each group and to contribute toward correlation of spectral changes and atomic structure.

Infrared radiation is absorbed by substances at those frequencies which coincide with certain vibrations of the atoms, molecules, and crystal structure of the absorbing substance. Fundamental vibration absorptions may be accompanied by overtone absorption bands which are multiples of fundamentals, and by combination bands which are sums or differences between fundamental frequencies. Details of the origin of infrared absorption spectra are discussed in textbooks on spectroscopy, and are referred to by prior articles on the infrared absorption of minerals (Keller and Pickett, 1949, 1950; Adler *et al.*, 1950; Hunt *et al.*, 1950).

The spectrograms reported in this paper were obtained by irradiating the mineral sample with a sequence of monochromatic frequencies of infrared radiation and measuring the percent of transmission at a large number of closely spaced wave-length intervals in the region being considered, and then plotting a curve of the percents of transmitted radiation as a function of the wave length or frequency. The instrument used in this study was a Beckman Model IR-2 infrared spectrophotometer.

The sample was prepared by first grinding about one gram of the mineral in a mechanically driven mullite mortar and pestle for a period of one to twelve hours. The powder was dispersed in distilled water and fractionated by sedimentation to obtain particles less than 2 microns in effective diameter. The suspension was concentrated by centrifuging, and the sample was washed several times with *n*-propyl alcohol to remove the water adsorbed during sedimentation.

Usually the water is removed satisfactorily, but it is difficult to remove all the adsorbed water from some few minerals, and as a consequence several spectra indicate absorption in the hydroxyl (2.7-3.1 microns) and molecular water (6.0-6.2) regions for minerals not containing these in their formulas. In those cases, additional samples of the minerals were prepared by dry grinding each mineral to a very fine powder, mulling the same powder in Nujol (heavy mineral oil) or Perfluoro Lube Oil and placing a smear of the paste on the same plate of the instrument. The spectrum of this smear was then scanned through the 2.7-3.1 and 6.0-6.2 microns regions. It was found that invariably the spectra of these few erals containing neither hydroxyl ions nor water molecules showed either no absorption or considerably less absorption in these regions than the spectra obtained from samples fractionated in water. In comparing spectra of minerals containing molecular water or hydroxyl ions in the lattice, scarcely any significant differences were noted in the spectra from samples prepared by the two methods.

Pfund (1934) found that fragmental particles cause scattering in the infrared when the wave length of radiation and the diameter of the particles are approximately equal. Scattering causes a decrease in transmission and degree of resolution. These effects have been reduced in this study by the use

of particles with diameters that are less than the shortest wave length of radiation employed.

MINERAL SPECTRA

The infrared absorption spectra of some of the more common carbonate, sulphate, phosphate (Biggs, 1950), and silicate (Spotts, 1951) minerals were determined. The minerals were chosen so that (a) the various structural classes within these groups were represented, and (b) the effect of changing either the cation or the anion could be evaluated when the other was held constant. The spectra have been grouped and discussed under structural classes in order to correlate insofar as possible the atomic structure and spectral characteristics. Tabulated locations of absorption maxima are given in subjoined tables, as are copies of the graphs of the spectra.

CARBONATE SPECTRA

The spectra of some of the common carbonates are catalogued in table 1 and figures 1-11. The carbonates include two isomorphous groups, the rhombohedral calcite group including those carbonates containing cations smaller than calcium, and the orthorhombic aragonite group including those with cations larger than calcium. Magnesite, siderite, rhodochrosite, calcite, and dolomite of the calcite group, and aragonite, strontianite, witherite, and cerussite of the aragonite group have been examined. The two groups are somewhat similar in structure but, in calcite, the CO_3 group is placed so that each oxygen atom "touches" two calcium atoms, whereas in aragonite each oxygen atom "touches" three calcium atoms. Slight structural differences arise from the different cations and their varying sizes.

Similarities in the structure of the carbonates are reflected in the similarities of their spectra, which are evident upon comparison of their graphs. In the following discussion, space will not be taken to repeat all the individual absorption peaks itemized in the tables or graphs except to show trends or to indicate points of special significance. Data in the tables and graphs will suffice for mineral determinative uses.

Examination of the graphs of plotted spectra show that all of the carbonates possess two prominent absorption peaks, one between 6.6 and 7.1 microns and one between 11.2 and

TABLE 1

CARBONATE MINERAL DATA AND LOCATIONS OF INFRARED ABSORPTION BANDS

MINERAL	COMPOSITION	DIVALENT CATION RADIUS	CATION MASS	WAVE LENGTH OF ABSORPTION MAXIMA IN MICRONS			
HEXAGONAL SYSTEM				1	2	3	4
1. Magnesite	$MgCO_3$	0.78	24.32	6.78	11.27	11.70	13.40
2. Siderite	$FeCO_3$	0.83	55.85	7.02	11.55	....	....
3. Rhodochrosite	$MnCO_3$	0.91	54.93	7.1	11.52	13.78	14.0
4. Calcite	$CaCO_3$	1.06	40.08	7.0	11.43	14.05	
				6.80	11.33		
5. Dolomite	$CaMg(CO_3)_2$	...	....	7.01	11.70	13.67	14.01
ORTHORHOMBIC SYSTEM							
6. Aragonite	$CaCO_3$	1.06	40.08	6.66	11.64	14.12	14.41
7. Strontianite	$SrCO_3$	1.27	87.63	6.80	11.65	....	....
8. Witherite	$BaCO_3$	1.43	137.6	6.89	11.67	....	....
9. Cerussite	$PbCO_3$	1.32	207.21	7.25	9.52	11.94	14.78
MIXTURES							
10. Aragonite +	$CaCO_3$			6.77	11.43		
Calcite				7.04	11.64		
11. Calcite +	$CaCO_3+MgCO_3$			6.80	11.35	11.68	13.45
Magnesite					11.22		14.0

MINERAL LOCALITIES

1. Magnesite, Snarum, Norway.
2. Siderite, Roxbury, Conn.
3. Rhodochrosite, Central City, Colo.
4. Calcite, Chihuahua.
5. Dolomite, Joplin, Mo.
6. Aragonite, Tolina Springs, Calif.
7. Strontianite, Hamm, Westphalia.
8. Witherite, locality unknown.
9. Cerussite, Coeur d'Alene, Idaho.
10. Mixture of Nos. 4 and 6.
11. Mixture of Nos. 4 and 1.

11.44 microns, which are due to absorption by ionic carbonate (CO_3).

It is obvious from a comparison of the infrared absorption spectra of the carbonates that the spectra are dominated by the anion. The exact location of the two main absorption bonds in the spectra may be affected by the cation present. In the spectra of the aragonite group the two major absorption peaks are spaced slightly more closely as the masses of the divalent cations increase (from calcium to lead). This observation has also been made by Adler *et al.* (1950). The effect of the cation on the position of the bands in the spectra of the calcite group is evidently not regular or simple.

Mixtures of carbonates: calcite and aragonite, and calcite and magnesite were measured in the infrared (figs. 10 and 11). The curves contain all the peaks that are present in the spectrum of each individual constituent of the mixture. Interference occurs when the various constituents have peaks that occur at approximately the same wave lengths, and thus the infrared absorption spectrum of a mixture of minerals comprises a curve that is a superposition of the spectra of all the mineral components of the mixture. Some difference between the spectrum of the calcite-magnesite mixture and the dolomite spectrum might be expected since dolomite is a distinct mineral species and the other was mixed mechanically.

The sulphate minerals which have been investigated in this study are barite, celestite, anhydrite, and selenite, repeating the cations of the carbonates. The spectra are illustrated in figures 12-15, and numerical data compiled in table 2.

The sulphate mineral spectra are characterized by a broad absorption band extending from about 7.5 to 10.5 microns in which minor absorption maxima occur. The sulfate ion $(SO_4)^{--}$ in solutions was found to absorb strongly from about 8.8 to 9.2 microns (American Cyanamid Lab., 1950). The two barite type minerals, barite and celestite, contain independent SO_4 groups with the cation surrounded by twelve oxygen atoms; their respective spectra are very similar. The main absorption band of barite is resolved into three absorption peaks located at 8.41, 8.92, and 9.21 microns, and that of celestite at 8.29, 8.84, and 9.26 microns.

The structure of anhydrite is different from that of the barite type, the difference being ascribed to the smaller size

TABLE 2

SULPHATE MINERAL DATA AND LOCATIONS OF THEIR INFRARED ABSORPTION BANDS

MINERAL	COMPOSITION	DIVALENT CATION RADIUS	CATION MASS	WAVE LENGTH OF ABSORPTION MAXIMA IN MICRONS				
				1	2	3	4	5
12. Barite	$BaSO_4$	1.43	137.46	8.41	8.92	9.21	10.22	
13. Celestite	$SrSO_4$	1.27	87.36	8.29	8.84	9.26	10.10	
14. Anhydrite	$CaSO_4$	1.06	40.08	6.19	6.90	8.65	10.90	14.26
15. Selenite	$CaSO_4 \cdot 2H_2O$	1.06	40.08	2.89	3.02	3.50	6.23	8.83

MINERAL LOCALITIES

12. Barite, Missouri.
13. Celestite, locality unknown.
14. Anhydrite, locality unknown.
15. Selenite, Wayne County, Utah.

of the calcium ion (Bragg, 1937). This structural difference is reflected also in the anhydrite spectrum (fig. 14). The main absorption band which is characteristic of the sulphate spectrum has only one peak which is located at 8.6 microns. Some hydration occurred during sample preparation of the anhydrite as indicated by absorptions located at 2.90 and 6.19 microns, which are known to be due to hydroxyl groups and molecular water, respectively. The spectrum of selenite (fig. 15) shows hydration to a greater extent by more pronounced absorption in the bonded hydroxyl region; absorption maxima occur at 2.98 and 3.02 microns. The presence of molecular water is shown by the band centered at 6.20 microns. The main absorption band in the selenite spectrum is identical in configuration and position with that of the anhydrite spectrum, indicating similar relationships within the anion groups of the two minerals.

PHOSPHATE SPECTRA

The spectrum of amblygonite and apatite are shown in figures 16 and 17. The band system in the phosphate spectra is similar to that of the sulphates, possibly due to similar structure of the tetrahedral anion groups of SO_4 in the sulphates and PO_4 in the phosphates. The numerical data are in table 3.

The apatite spectrum displays peaks within the main band at 9.09 and 9.65 microns and a slighter absorption at 10.42 microns. The major absorption band in the amblygonite curve is considerably broader and shows absorption maxima at 8.37, 9.01, 9.70, and 9.95 microns. Absorption also occurs at 12.22 microns, but an analogous band was not detected in the apatite spectrum.

ANION AND CATION INFLUENCES ON THE SPECTRA

It is possible to evaluate the relative effects of the anions and cations on the infrared absorption spectra by comparing the calcium-containing minerals, for example, as against the carbonate minerals. Calcium carbonate, calcium sulphate, calcium phosphate, and calcium silicate have little in common compared to the members within the carbonate, the sulphate, or other anionic groups. Just as the crystal architecture is dominated by the large oxygen-containing anions, so are the

TABLE 8

PHOSPHATE MINERAL DATA AND LOCATIONS OF THEIR INFRARED ABSORPTION BANDS

MINERAL	COMPOSITION	WAVE LENGTH OF ABSORPTION MAXIMA IN MICRONS						
		1	2	3	4	5	6	7
16. Amblygonite	$LiAlFPO_4 \cdot H_2O$	2.95	6.1	8.37	9.01	9.70	9.95	12.22
17. Apatite	$Ca_5F(PO_4)_3$	9.09	9.65	10.42				

MINERAL LOCALITIES

16. Amblygonite, South Dakota.

17. Apatite, locality unknown.

absorption spectra dominated by the bonds between the oxygen and the element which characterizes the anion. The cations exert only a secondary influence.

From another viewpoint, the infrared absorption spectra owe their origin to the covalent bonding which is accentuated in the anions.

SILICATE SPECTRA

The silicate minerals exhibit a much wider variety of structural types than any of the other groups discussed and, analogously, the spectra of the silicates show a greater variety of absorption patterns and intensities. The silicates are classified on the basis of linkage of the silica tetrahedra, using the terminology of Winchell and Winchell (1951), as follows: the neosilicates (separate SiO_4 groups); the sorosilicates (Si_2O_7 groups of two linked tetrahedra); the cyclosilicates (rings of linked tetrahedra); the inosilicates (pyroxene and amphibole chains); the phyllosilicates (sheet structure); and the tectosilicates (framework structure). One or more representatives of each group were run but these few are obviously far from being exhaustive of the groups. Tabulated locations of absorptions are given in table 4.

Nesosilicate spectra.—Olivine and willemite, two representatives of the nesosilicates, show close relationship in their spectra. Their absorption patterns (figs. 18 and 19) are similar in that there is a large absorption band centered at 11.1-11.2 microns and that this band contains four absorption maxima which are comparable in relative intensities. It is noticeable that the olivine spectrum is spread over a wider range of wave lengths than those in willemite. The configuration of the curves compares favorably with the garnet spectra (a nesosilicate) determined by Adler *et al.* (1950), although the absorption peaks in the olivine and willemite spectra occur at shorter wave lengths.

An interesting variation of the nesosilicate structure is present in the Al_2SiO_5 minerals, in which the SiO_4 tetrahedra are independent, except that pairs of tetrahedra are adjacent in kyanite. The Al_2SiO_5 minerals are of further interest because of their contrasting structures, and because they lack the univalent and divalent ions which occur commonly in the silicate minerals. All three are built of chains of octahedral

TABLE 4

DATA ON SILICATE MINERALS AND LOCATIONS OF THEIR INFRARED ABSORPTION BANDS

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13
NESOSILICATES													
18. Olivine	9.95	10.80	11.20	11.90	....	....	....	....	....	....	....	....	....
19. Willemite	9.25	10.27	10.75	11.07	11.50	....	....	....	....	....	....	....	....
20. Kyanite	2.95	6.15	8.60	8.95	11.15	12.51	12.85	13.75	14.06	....	....	....	....
21. Andalusite	2.19	3.0	6.15	9.45	9.85	10.75	12.0	12.49	13.4	13.9	14.40	....	....
22. Sillimanite	2.95	6.10	6.95	8.47	10.45	10.80	11.2	12.15	13.37	14.3	....	....	....
23. Staurolite	2.90	6.10	8.60	9.02	10.10	11.70	12.50	12.85	13.80	14.60	....	....	....
24. Idocrase	2.82	9.88	10.40	10.90	11.32	12.40	....	....	....	....	....	....	....
SOROSILICATES													
25. Hemimorphite	3.00	6.20	9.22	10.75	11.60	....	....	....	....	....	....	....	....
26. Melilite	9.50	9.75	10.18	10.63	11.60	....	....	....	....	....	....	....	....
CYCLOSILICATES													
27. Beryl	8.26	9.85	10.44	12.32	13.40	14.65	....	....	....	....	....	....	....
28. Benitoite	9.60	10.65	13.05	....	....	....	....	....	....	....	....	....	....
29. Cordierite	8.50	8.75	9.81	10.40	11.00	12.98	13.40	14.30	....	....	....	....	....
30. Tourmaline	6.90	7.45	7.68	8.42	9.00	9.70	10.00	12.72	13.84	....	....	....	....
INOSILICATES													
31. Diopside	9.24	10.30	10.87	11.35	13.10	....	....	....	....	....	....	....	....
32. Hypersthene	9.32	9.92	10.62	11.15	11.55	13.80	14.50	....	....	....	....	....	....
33. Anthophyllite	2.95	6.16	9.03	9.75	10.18	11.00	12.80	13.24	14.10	....	....	....	....
34. Tremolite	3.00	6.20	6.92	8.52	9.10	9.45	9.78	10.37	10.68	11.06	12.00	13.45	14.70
PHYLLOSILICATES													
35. Muscovite	2.72	6.15	9.75	10.86	11.92	12.50	13.36	14.60	....	....	....	....	....
36. Biotite	2.78	6.18	9.94	12.84	14.14	....	....	....	....	....	....	....	....
39. Pyrophyllite	2.80	6.20	8.92	9.26	10.55	10.92	11.72	11.95	13.30	14.45	....	....	....
40. Talc	2.90	6.12	9.84	12.53	12.82	....	....	....	....	....	....	....	....
41. Prochlorite	2.84	2.95	6.12	9.94	13.30	....	....	....	....	....	....	....	....

TABLE 4 (Continued)

DATA ON SILICATE MINERALS AND LOCATIONS OF THEIR INFRARED ABSORPTION BANDS

	1	2	3	4	5	6	7	8	9	10	11	12	13
TECTOSILICATES													
42. Quartz	6.30	6.90	8.00	8.60	9.20	12.50	12.83	14.45	....	....	....	....	....
44. Orthoclase	8.80	9.62	12.95	13.80	....	....	....	....	....	....	....	....	....
45. Adularia	8.84	9.62	12.95	13.75	....	....	....	....	....	....	....	....	....
46. Microcline	8.75	9.85	12.95	13.72	....	....	....	....	....	....	....	....	....
48. Clevelandite	8.65	9.08	9.65	10.00	12.74	13.16	13.45	13.80	....	....	....	....	....
49. Analcime	2.85	6.14	9.04	9.60	11.30	13.60	....	....	....	....	....	....	....
50. Nepheline	9.22	10.00	14.27	14.00	....	....	....	....	....	....	....	....	....
51. Leucite	9.70	13.80	....	....	....	....	....	....	....	....	....	....	....
52. Heulandite	2.95	6.15	9.30	11.20	12.50	12.82	....	....	....	....	....	....	....
54. Natrolite	2.94	6.10	9.85	14.30	....	....	....	....	....	....	....	....	....

MINERAL LOCALITIES

18. Olivine, Jackson Co., N. C.
19. Willemite, probably Franklin, N. J.
20. Kyanite, Los Tablos, N. Mex.
21. Andalusite, Madera Co., Calif.
22. Sillimanite, Khasia, India.
23. Staurolite, Faunin Co., Ga.
24. Idocrase, Crestmore, Calif.
25. Hemimorphite, locality unknown.
26. Melilite, Iron Hill, Colo.
27. Beryl, locality unknown.
28. Benitoite, San Benito, Calif.
29. Cordierite, Kragero, Norway.
30. Tourmaline, locality unknown.
31. Diopside, locality unknown.
32. Hypersthene, locality unknown.
33. Anthophyllite, Bakersville, N. C.
34. Tremolite, locality unknown.
35. Muscovite, Jackson Co., N. C.
36. Biotite, locality unknown.
37. Muscovite flake, Jackson Co., N. C.
38. Biotite flake, locality unknown.
39. Pyrophyllite, Robbins, N. C.
40. Talc, St. Lawrence Co., N. Y.
41. Prochlorite, Chester, Vermont.
42. Quartz crystal, Arkansas.
43. Rose quartz, Custer, S. Dak.
44. Orthoclase, Crook Co., Wyo.
45. Orthoclase (Adularia), St. Gotthard, Switzerland.
46. Microcline (pink), Verona, Ontario.
47. Microcline (Amazonstone), Amelia Court House, Va.
48. Clevelandite, locality unknown.
49. Analcite, Table Mt. Golden, Colo.
50. Nepheline, locality unknown.
51. Senate, near Rome, Italy.
52. Heulandite, Iceland.
53. Diatomite, Quincy, Wash.
54. Natrolite, Medford, Oregon.

alumina, but in sillimanite the chains are linked by tetrahedral alumina and silica, in andalusite by five-fold coordinated alumina and tetrahedral silica, and in kyanite, by octahedral alumina and tetrahedral silica. Their infrared absorption spectra are shown in figures 20, 21, and 22.

Strong absorption begins in kyanite in the 8.0 to 8.5 micron region, where quartz also absorbs strongly, and continues until almost 12 microns, which resembles the phyllosilicate absorption. Two small absorption peaks at 12.5 and about 12.8 microns again bear some similarity to quartz. The andalusite curve is less sharply defined than that of kyanite which it resembles somewhat. Sillimanite absorbs strongly between 8.0 and 8.5 microns, like that part of the kyanite spectrum, after which the curve rises sharply. The absence of absorption between 8.8 and 10 microns is unusual for silicate minerals. One wonders if the unorthodox absorption by sillimanite may not be due to the effect of the tetrahedral alumina. Adler's curve for sillimanite similarly differed from those of kyanite and andalusite.

Staurolite may be thought of structurally as consisting of alternating layers of kyanite and iron hydroxide. The absorption spectrum of staurolite (fig. 23) closely resembles that of kyanite plus additional OH (from the hydroxide of iron) at about 2.8 microns.

Another variation of a nesosilicate is idocrase, which is built of independent SiO_4 groups and Si_2O_7 groups (as in the sorosilicates). The spectrum of idocrase is shown in figure 24. Its absorption bands are present at about the same wave lengths as in the garnet spectrum, but the shape of the curve is more closely related to that of melilite, which is discussed under the sorosilicates which follow.

Sorosilicate spectra.—The spectra of the sorosilicates, hemimorphite and melilite show good correlation (figs. 25 and 26). The absorptions are shifted toward shorter wave lengths with respect to the nesosilicate spectra. The broad absorption band in this group spans from 9.0 to 12.0 microns, and maximum absorption occurs at 10.64 microns in the melilite curve, and at 10.75 microns in the hemimorphite curve.

Cyclosilicate spectra.—The cyclosilicates which have been considered are benitoite with rings of three silica tetrahedra, beryl with rings of six tetrahedra, and cordierite and tour-

maline which are structurally related to beryl. The beryl and benitoite spectra are distinctly different, as illustrated in figures 27 and 28.

The cordierite spectrum (fig. 29) and the beryl spectrum show distinct similarities. Concerning the cordierite structure, Bragg (1937) wrote, "It is very probable that the orthorhombic crystal cordierite is closely related in structure to beryl." The two spectra obtained support this view.

The tourmaline curve (fig. 30) is much more complex than the other cyclosilicate spectra, and it shows general resemblance to that of beryl, possibly due to structural relationships. The spectrum of tourmaline possesses much more fine structure and spans a greater range in wave lengths. Structurally, the rings of six silica tetrahedra in tourmaline differ from the rings in beryl in that there is no equatorial plane of symmetry in the tourmaline ring, and this difference may be responsible, in part, for spectral dissimilarities.

Inosilicate spectra.—The inosilicates include the pyroxenes and amphiboles. The spectra of inosilicates determined in this study are shown in figures 31-34. The pyroxenes, diopside and hypersthene, are very similar in structural arrangements and their spectra show similarities that are in keeping with this. The main absorption peak extends from 8.5 to 12.0 microns and contains accessory minor absorption maxima.

Agreement in spectra is not quite so evident in the curves of the amphiboles, anthophyllite, and tremolite. The atomic arrangement of tremolite is more closely related to that of the pyroxenes than is anthophyllite, and its absorption pattern is more similar to the pyroxene curves.

Phyllosilicate spectra.—The absorption spectra of the phyllosilicates are shown in figures 35-41. All the spectra in this group are characterized by one main absorption band which is centered between 9.0 and 10.0 microns. Hydroxyl absorbs strongly at 2.75 microns in muscovite and other micas (Keller and Pickett, 1950; Tsuboi, 1950).

The pyrophyllite curve is somewhat more complex than the other phyllosilicate curves; part of this greater complexity is due to the presence of a small amount of quartz impurity as indicated by the absorptions at 12.50 and 12.84 microns.

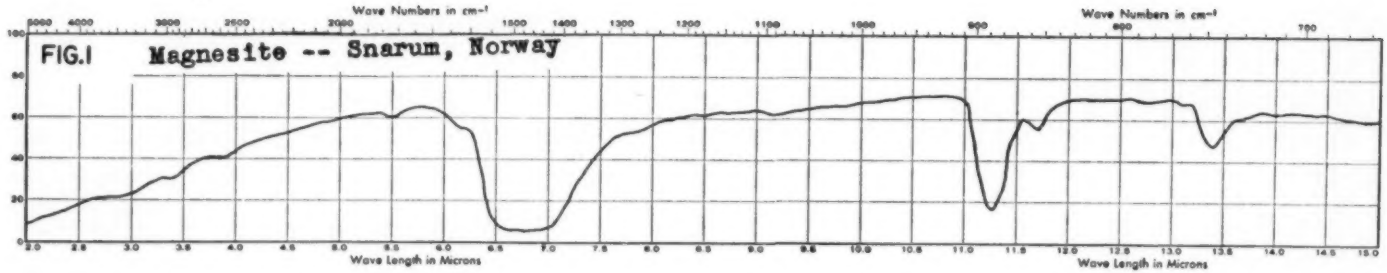


FIG.1 Magnesite -- Snarum, Norway
Wave Numbers in cm⁻¹
Wave Length in Microns

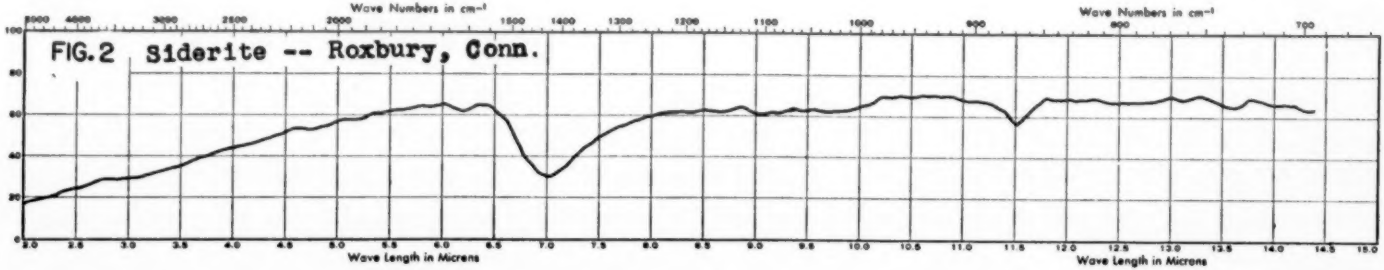


FIG.2 Siderite -- Roxbury, Conn.
Wave Numbers in cm⁻¹
Wave Length in Microns

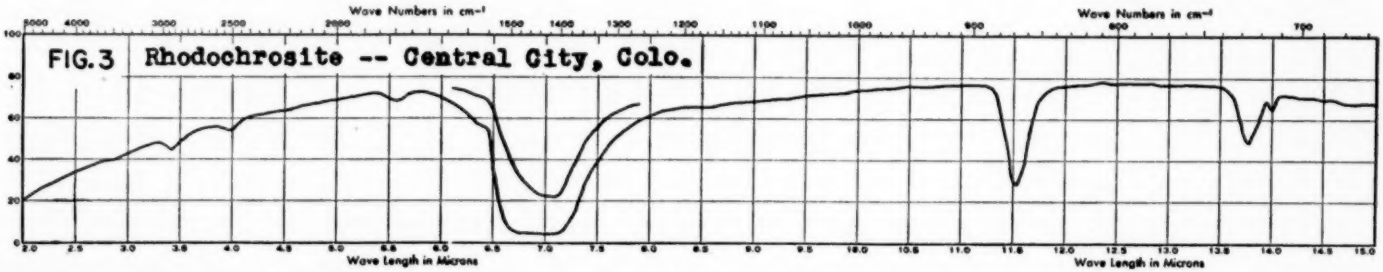


FIG.3 Rhodochrosite -- Central City, Colo.
Wave Numbers in cm⁻¹
Wave Length in Microns

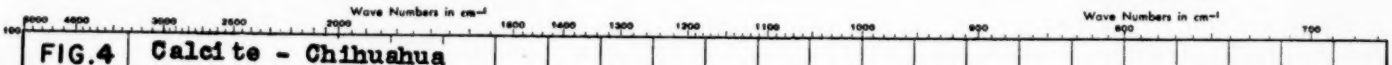


FIG.4 Calcite - Chihuahua
Wave Numbers in cm⁻¹
Wave Length in Microns

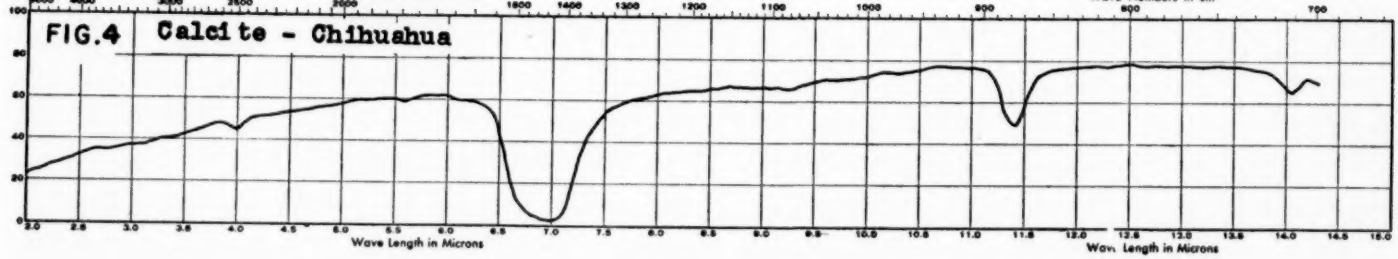


FIG.4 Calcite - Chihuahua
Wave Numbers in cm⁻¹
Wave Length in Microns

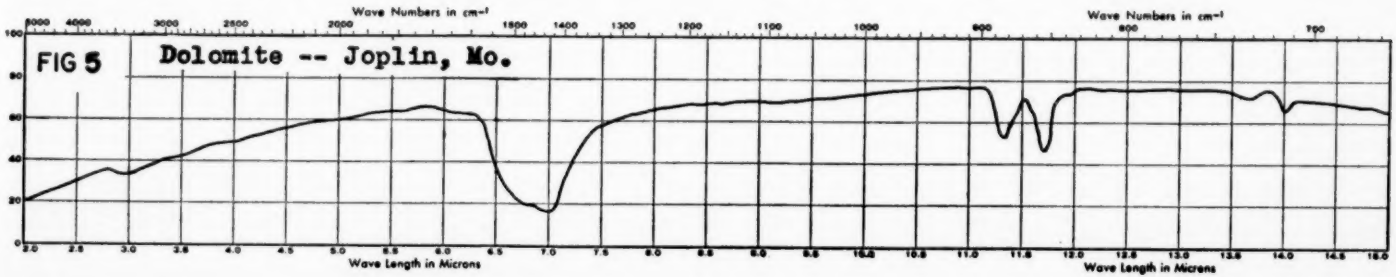


FIG 5 Dolomite -- Joplin, Mo.
Wave Numbers in cm⁻¹
Wave Length in Microns

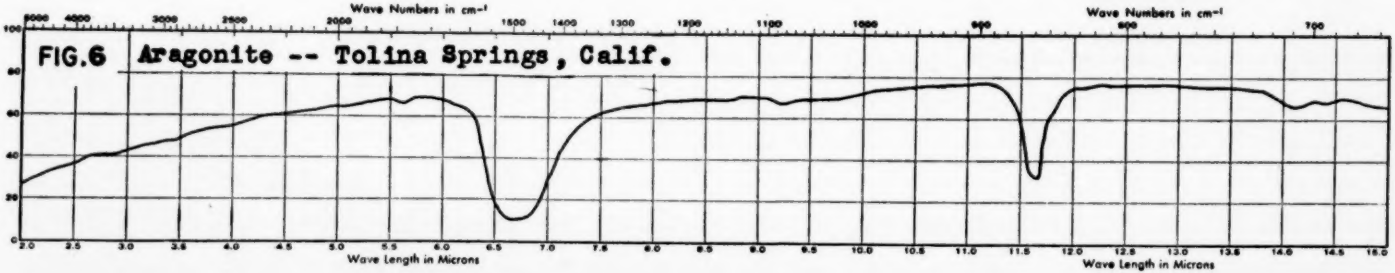


FIG.6 Aragonite -- Tolina Springs, Calif.
Wave Numbers in cm⁻¹
Wave Length in Microns

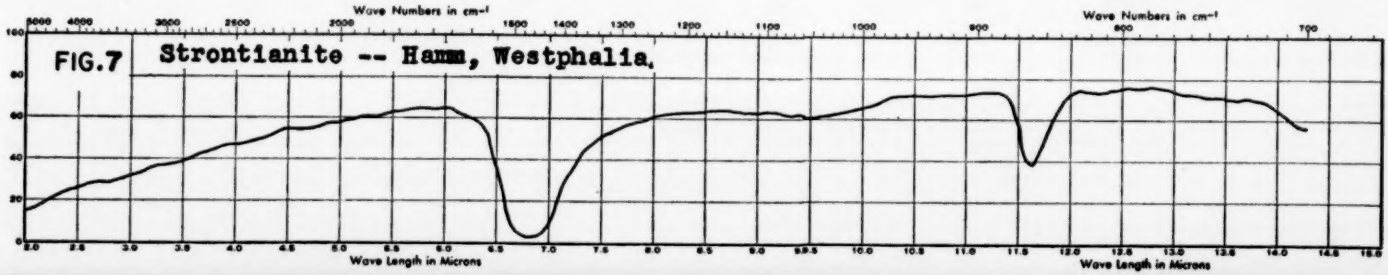


FIG.7 Strontianite -- Hamm, Westphalia.
Wave Numbers in cm⁻¹
Wave Length in Microns

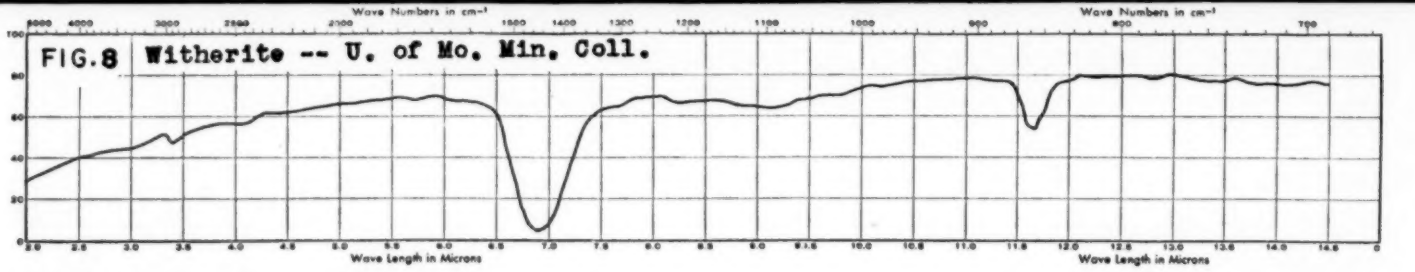


FIG.8 Witherite -- U. of Mo. Min. Coll.

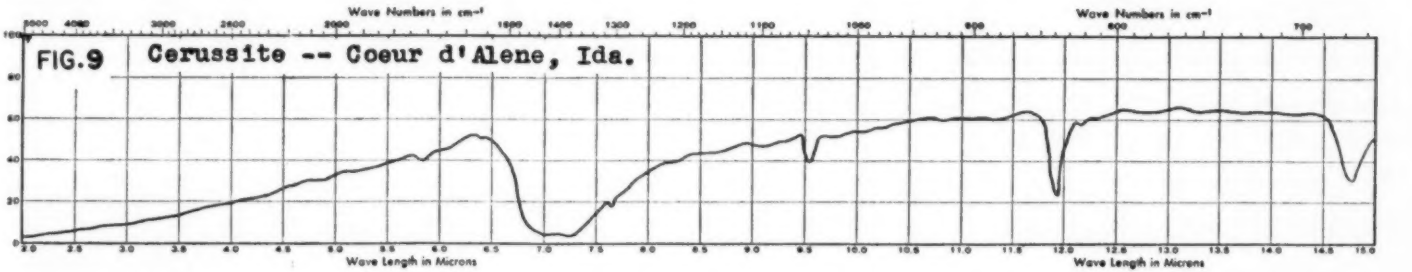


FIG.9 Cerussite -- Coeur d'Alene, Ida.

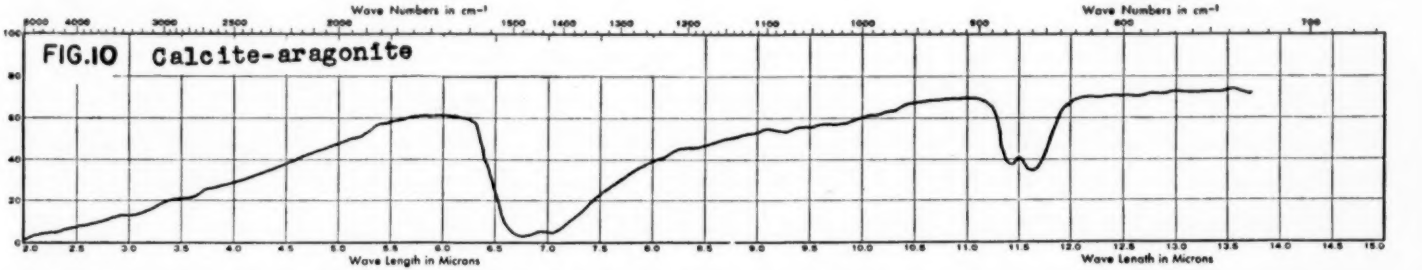


FIG.10 Calcite-aragonite

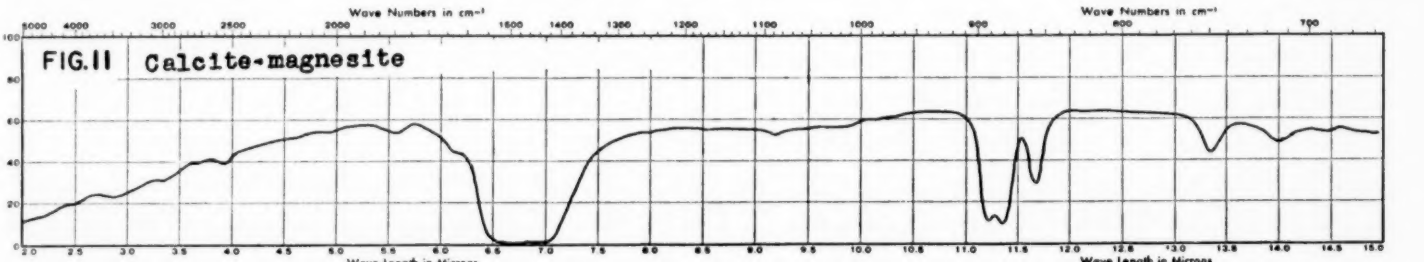


FIG.11 Calcite-magnesite

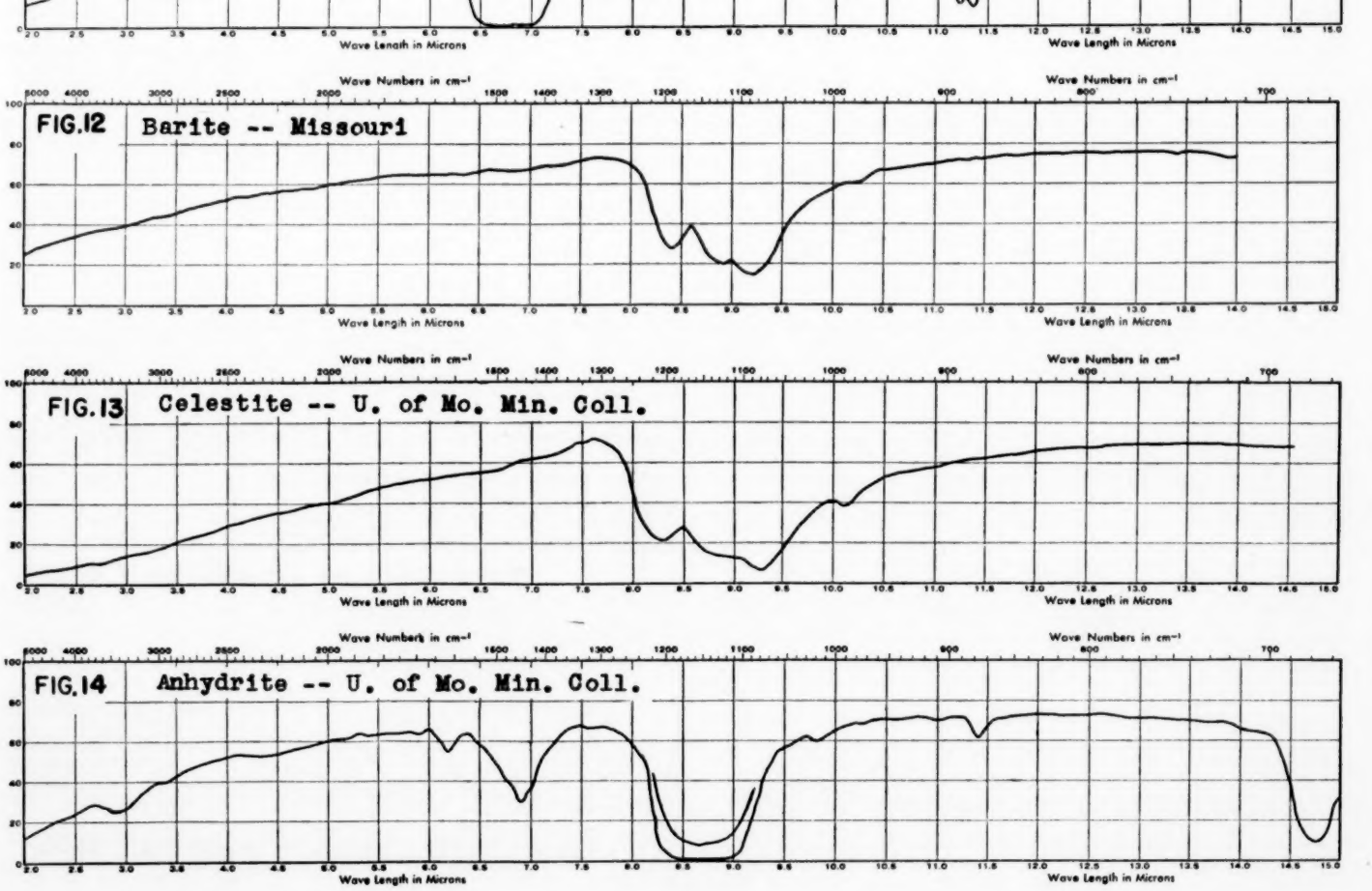


FIG.12 Barite -- Missouri
FIG.13 Celestite -- U. of Mo. Min. Coll.
FIG.14 Anhydrite -- U. of Mo. Min. Coll.
Wave Numbers in cm⁻¹
Wave Length in Microns

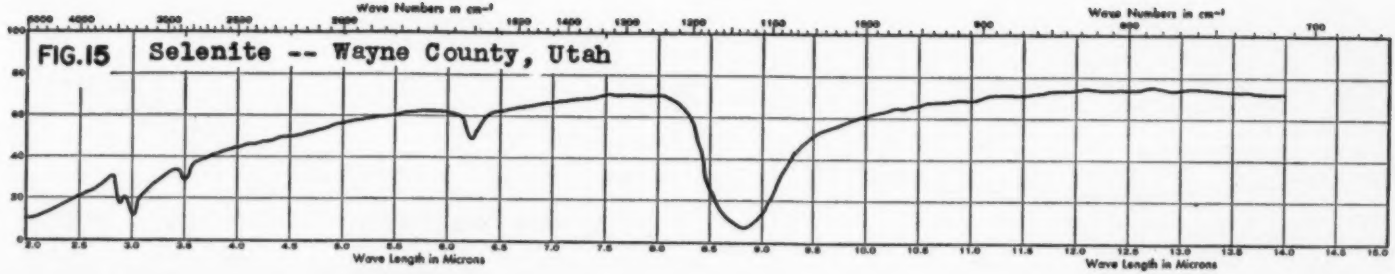


FIG.15 Selenite -- Wayne County, Utah
Wave Numbers in cm⁻¹
Wave Length in Microns

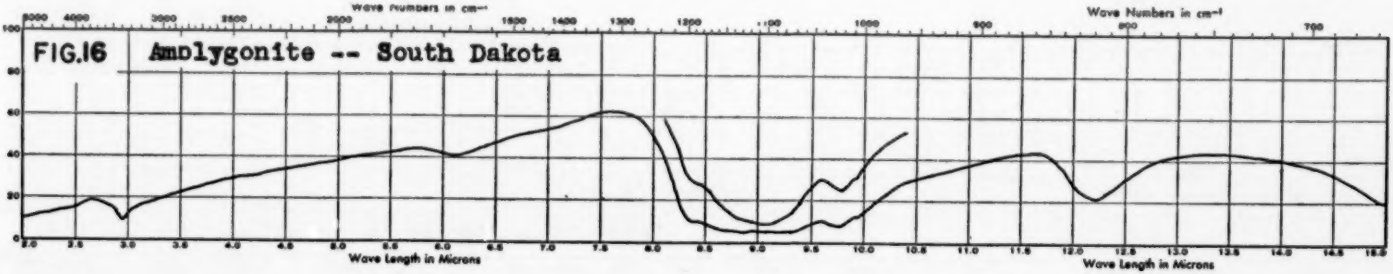


FIG.16 Amblygonite -- South Dakota
Wave Numbers in cm⁻¹
Wave Length in Microns

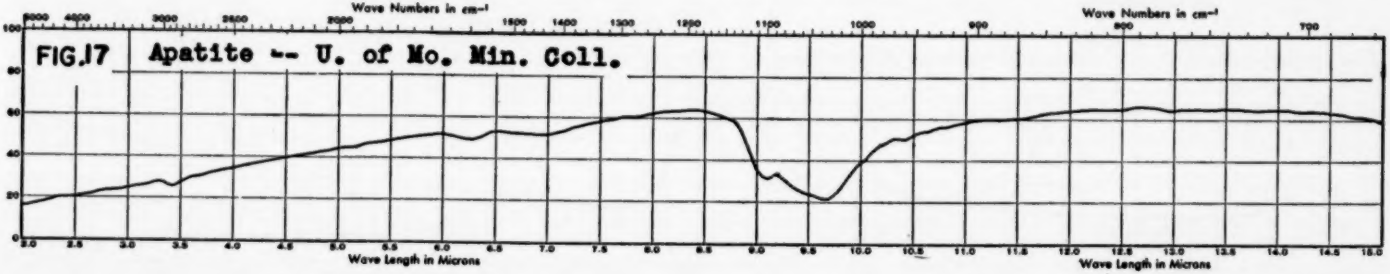
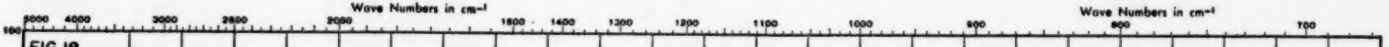


FIG.17 Apatite -- U. of Mo. Min. Coll.
Wave Numbers in cm⁻¹
Wave Length in Microns


Wave Numbers in cm⁻¹
Wave Length in Microns

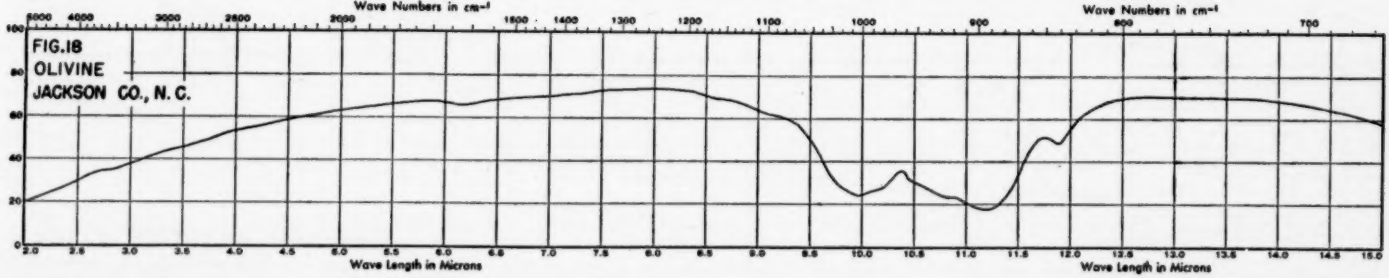


FIG.18
OLIVINE
JACKSON CO., N. C.
Wave Numbers in cm⁻¹
Wave Length in Microns

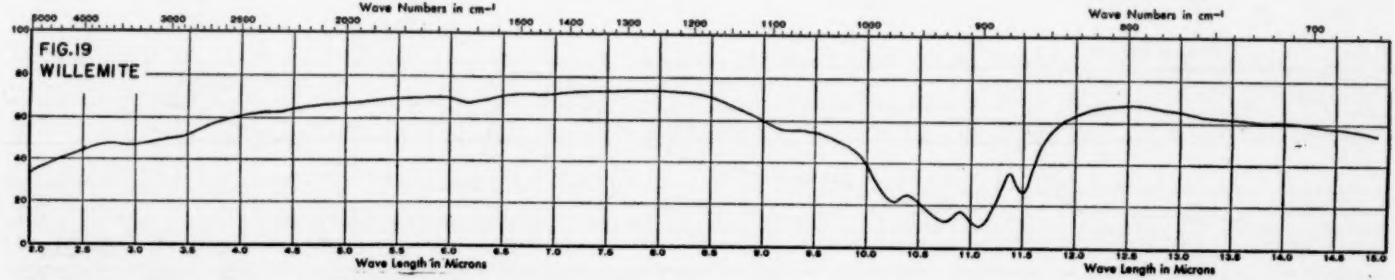


FIG.19
WILLEMITE
Wave Numbers in cm⁻¹
Wave Length in Microns

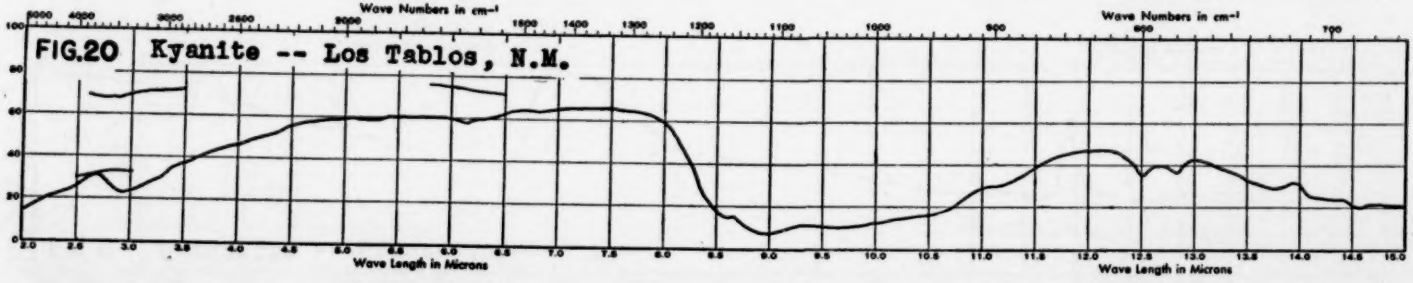


FIG.20
Kyanite -- Los Tablos, N.M.
Wave Numbers in cm⁻¹
Wave Length in Microns

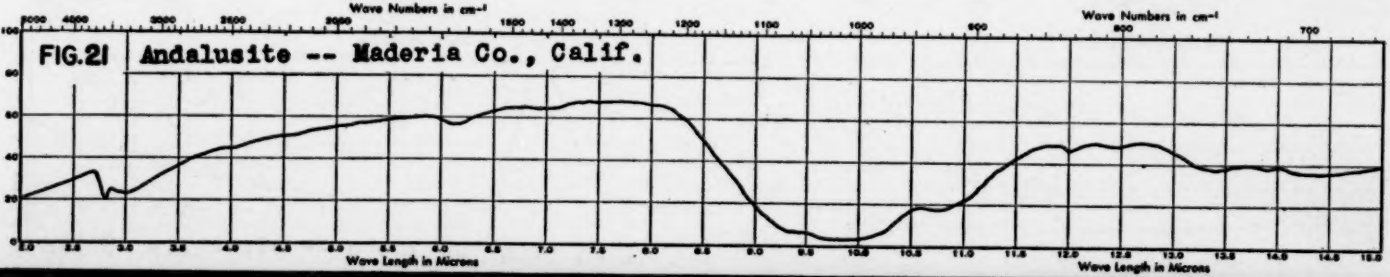


FIG.21
Andalusite -- Maderia Co., Calif.
Wave Numbers in cm⁻¹
Wave Length in Microns

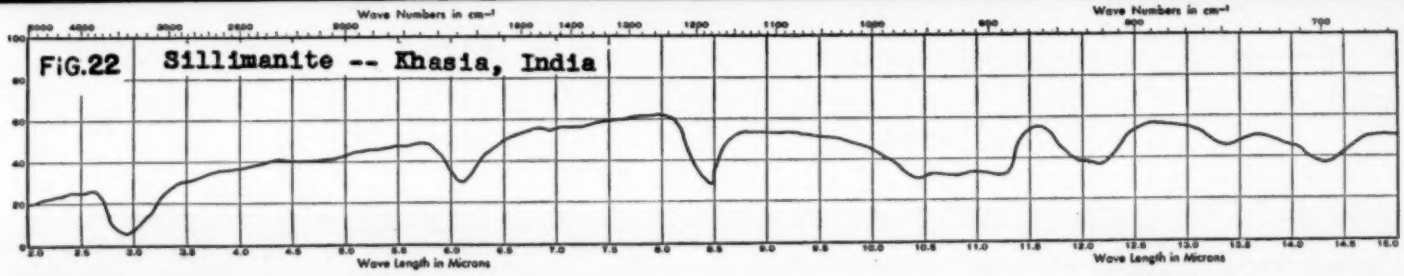


FiG.22 Sillimanite -- Khasia, India
Wave Numbers in cm⁻¹
Wave Length in Microns

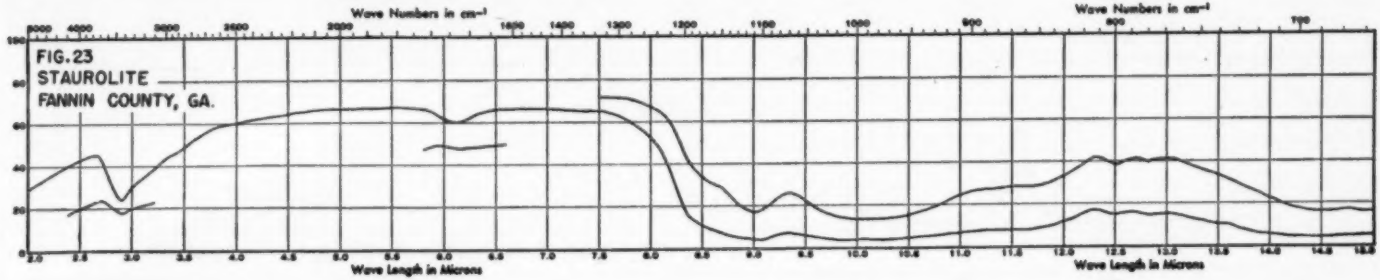


FIG.23
STAUROLITE
FANNIN COUNTY, GA.
Wave Numbers in cm⁻¹
Wave Length in Microns

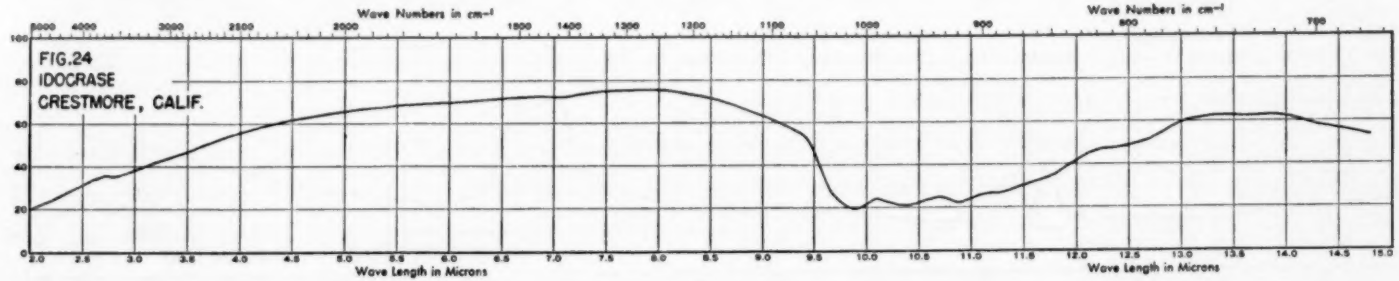


FIG.24
IDOCRASE
CRESTMORE, CALIF.
Wave Numbers in cm⁻¹
Wave Length in Microns

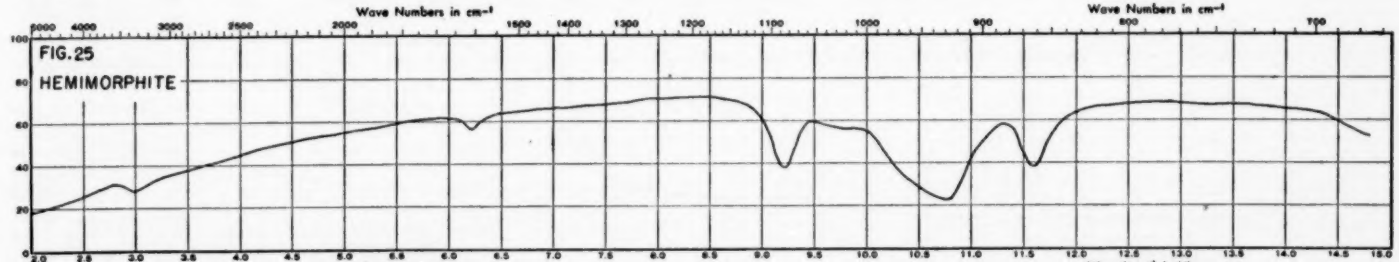
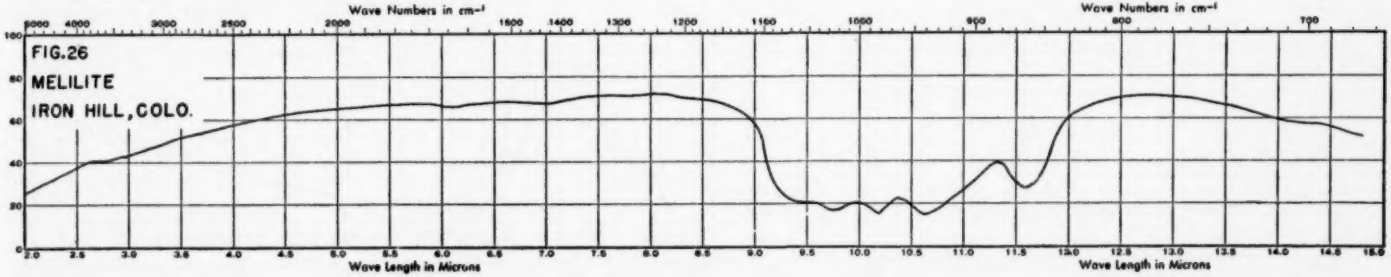
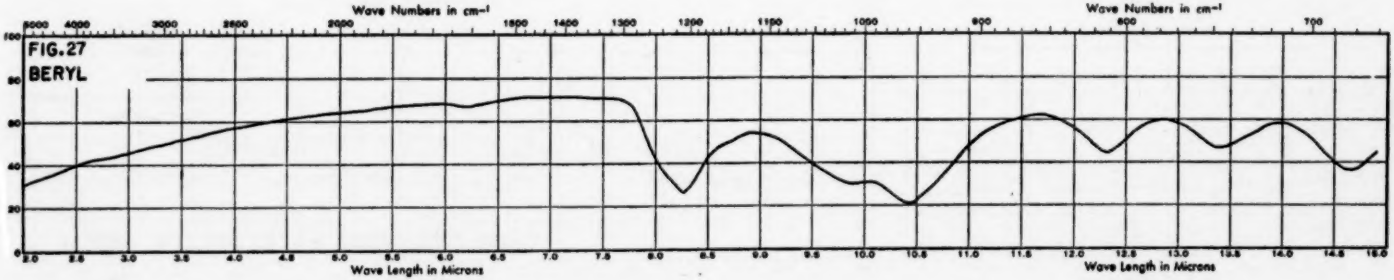
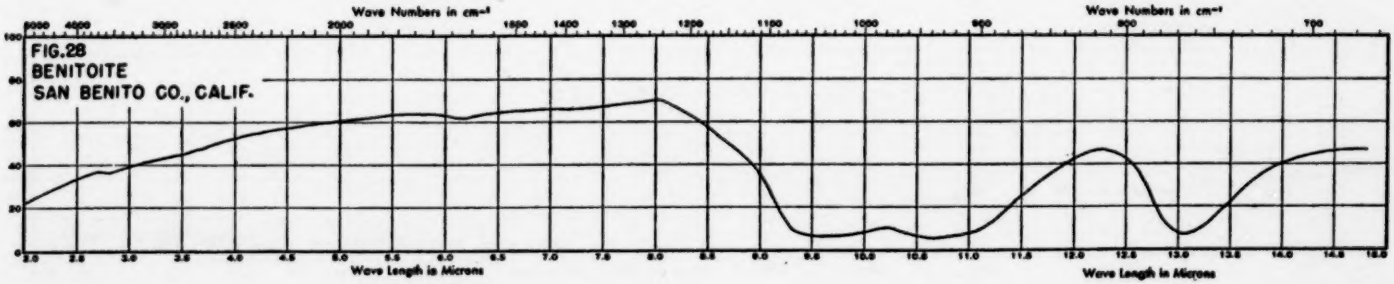
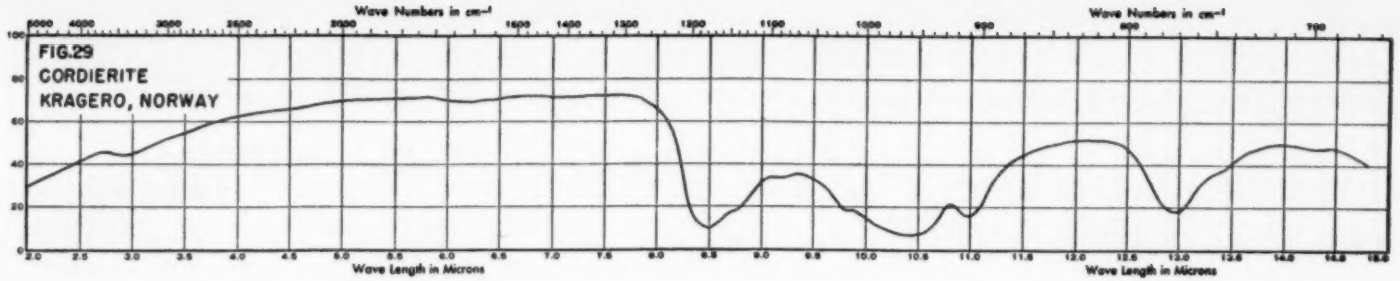


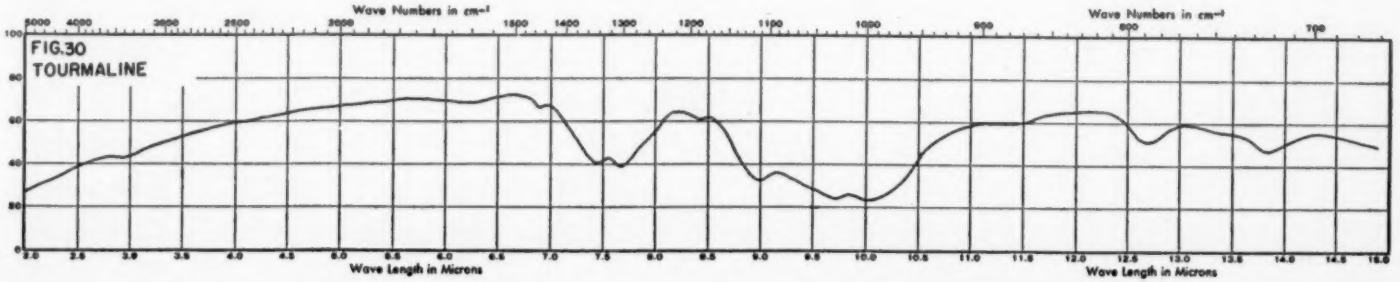
FIG.25
HEMIMORPHITE
Wave Numbers in cm⁻¹
Wave Length in Microns

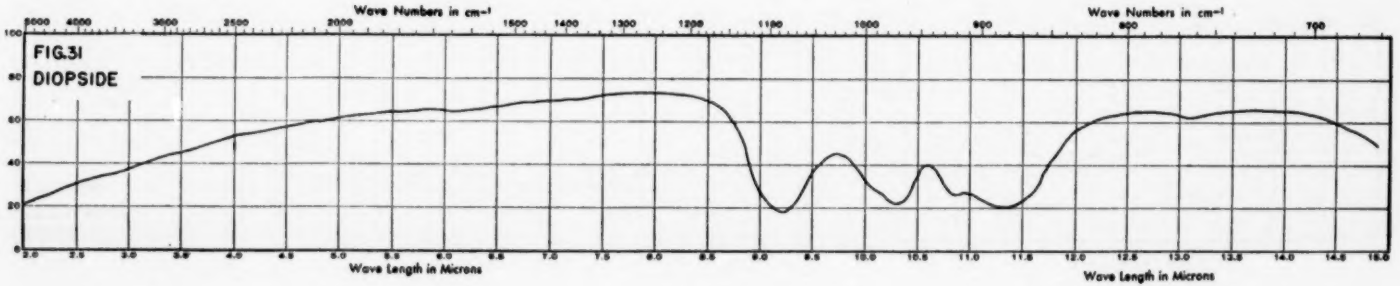

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.26
MELILITE
IRON HILL, COLO.

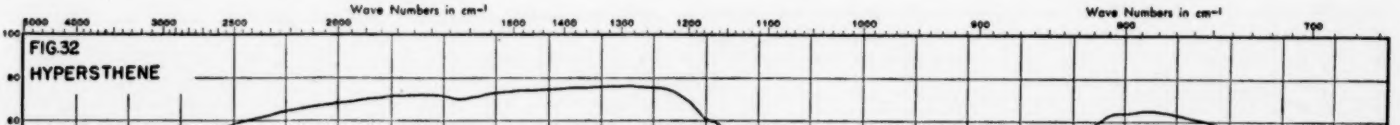

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.27
BERYL

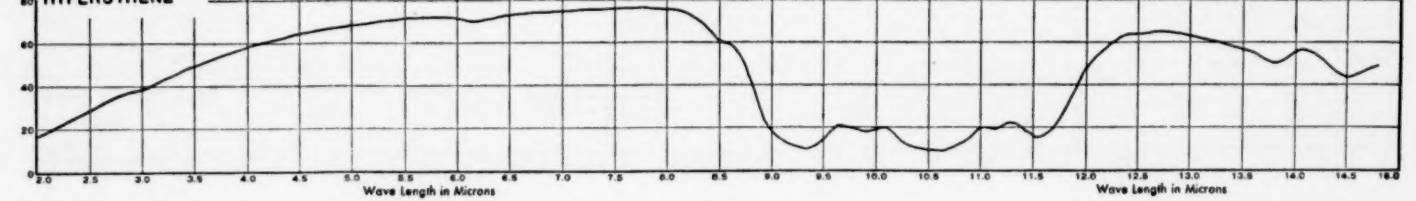

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.28
BENITOITE
SAN BENITO CO., CALIF.


Wave Numbers in cm⁻¹
FIG.29
CORDIERITE
KRAGERO, NORWAY
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.30
TOURMALINE
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.31
DIOPSIDE
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.32
HYPERSTHENE


Wave Numbers in cm⁻¹
Wave Length in Microns

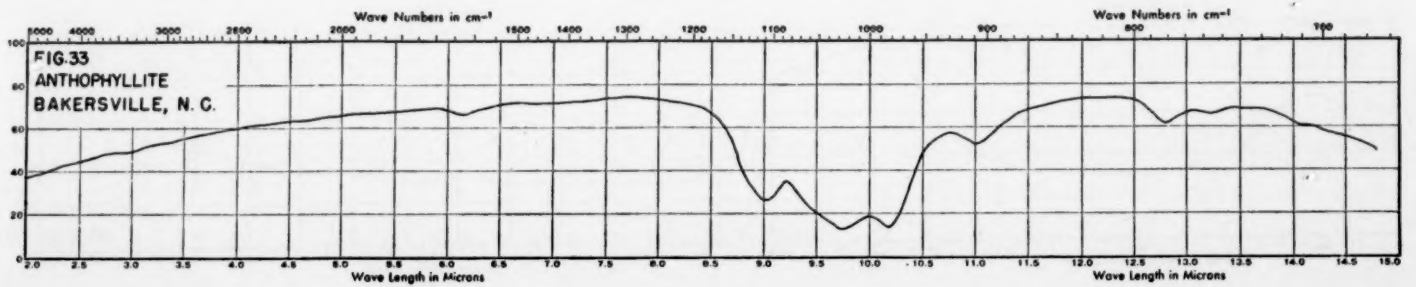


FIG.33
ANTHOPHYLLITE
BAKERSVILLE, N. C.
Wave Numbers in cm⁻¹
Wave Length in Microns

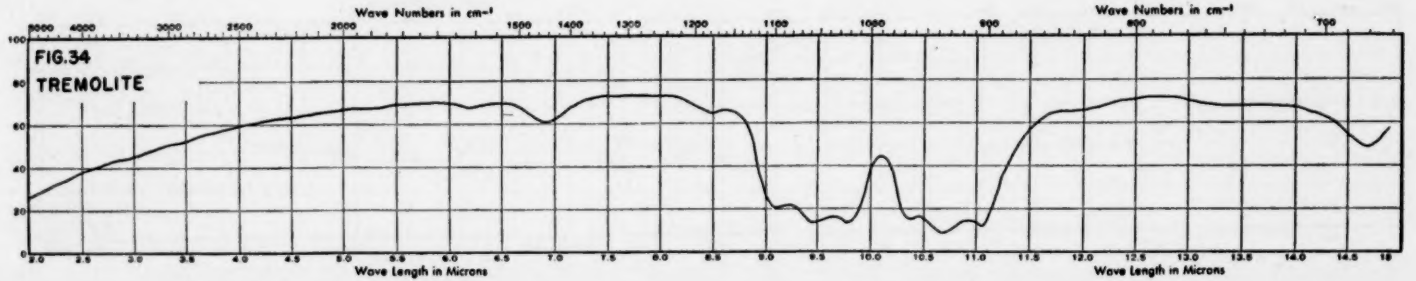


FIG.34
TREMOLITE
Wave Numbers in cm⁻¹
Wave Length in Microns

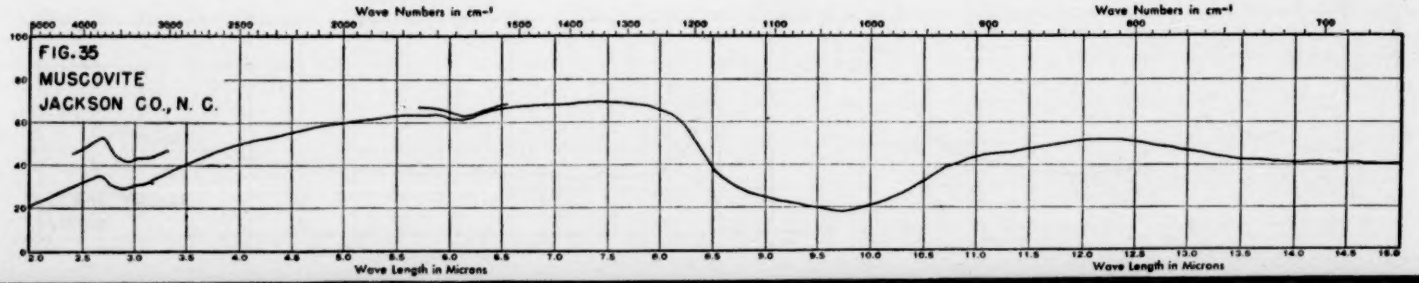


FIG.35
MUSCOVITE
JACKSON CO., N. C.
Wave Numbers in cm⁻¹
Wave Length in Microns

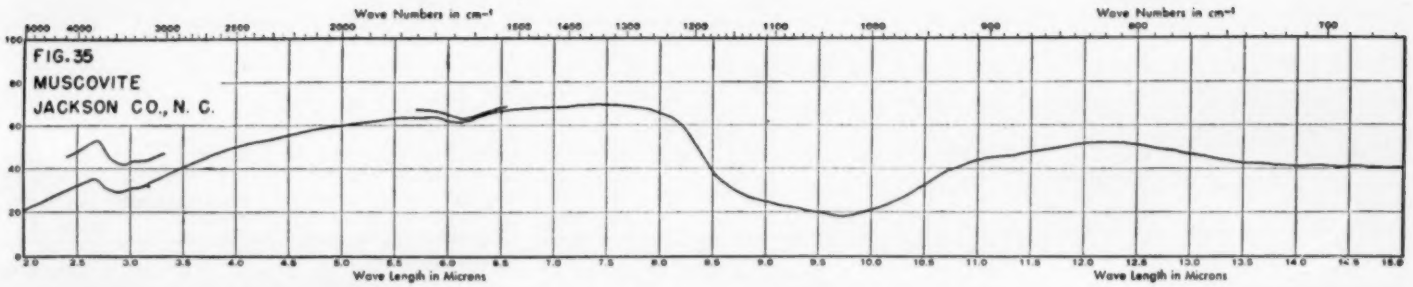


FIG. 35
MUSCOVITE
JACKSON CO., N. C.
Wave Numbers in cm⁻¹
Wave Length in Microns

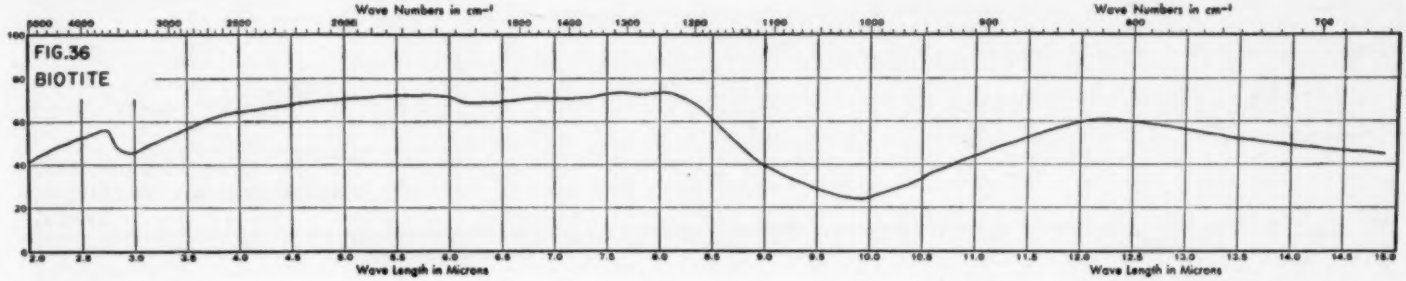


FIG. 36
BIOTITE
Wave Numbers in cm⁻¹
Wave Length in Microns

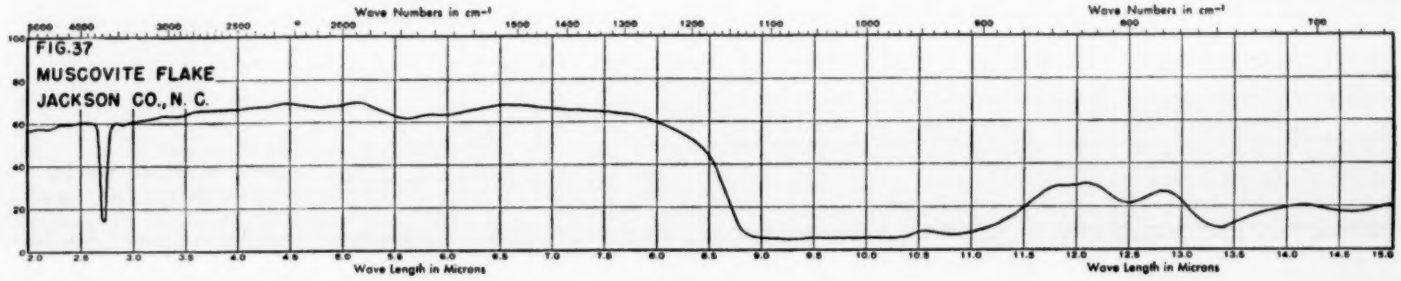


FIG. 37
MUSCOVITE FLAKE
JACKSON CO., N. C.
Wave Numbers in cm⁻¹
Wave Length in Microns

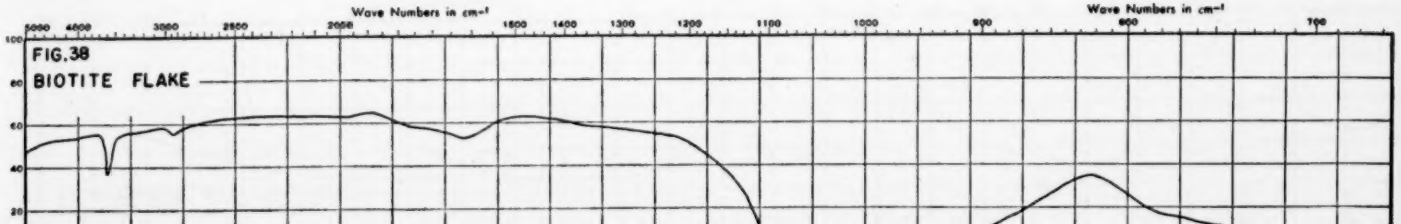
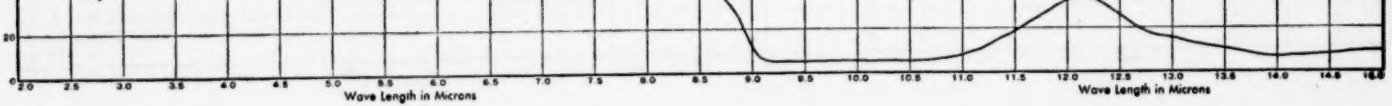
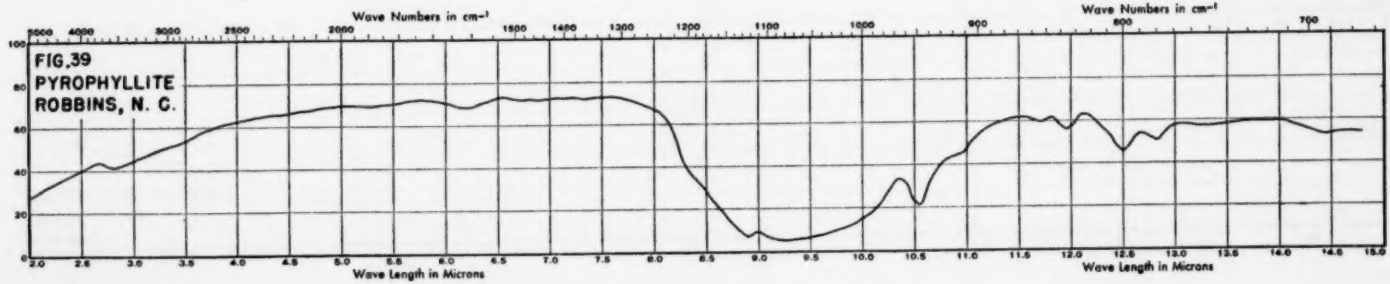
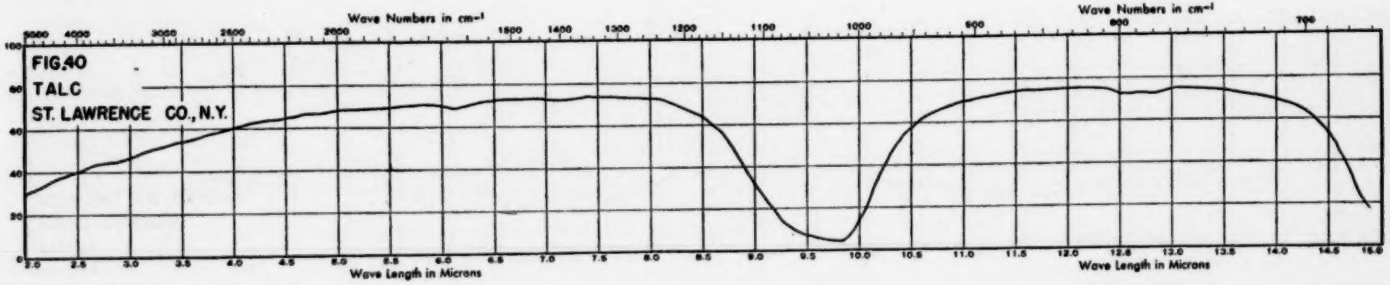
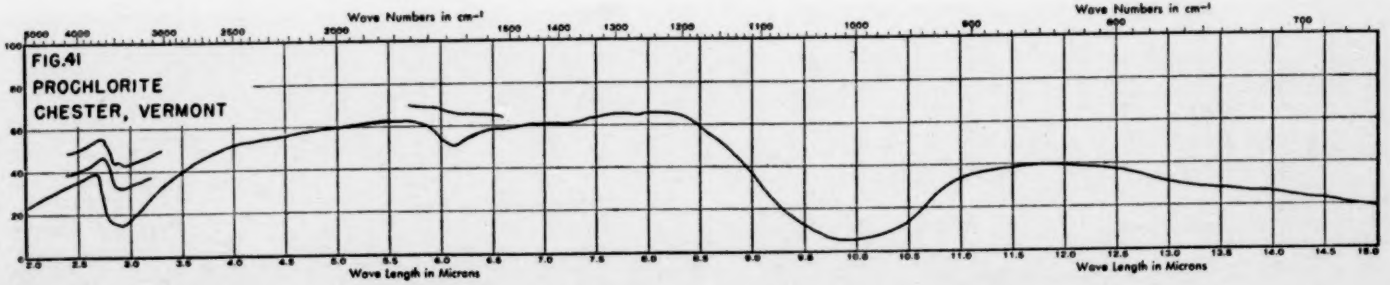


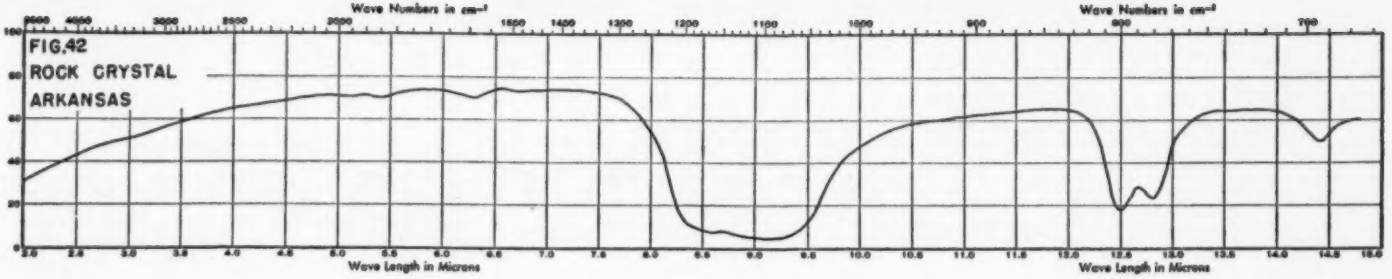
FIG. 38
BIOTITE FLAKE
Wave Numbers in cm⁻¹
Wave Length in Microns

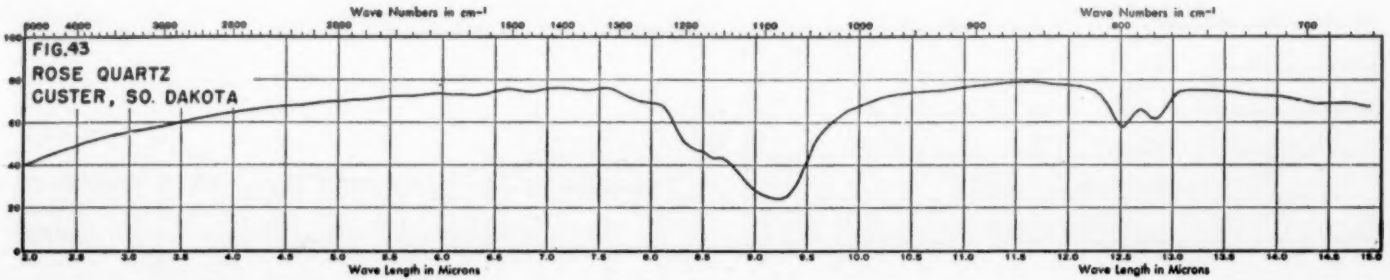

Wave Length in Microns
Wave Length in Microns

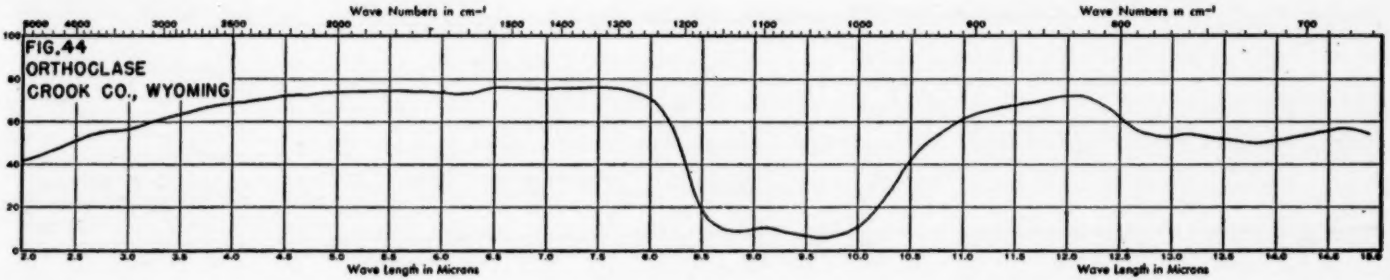

Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.39
PYROPHYLLITE
ROBBINS, N. C.
Wave Length in Microns
Wave Length in Microns

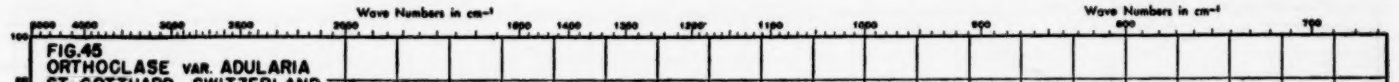

Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.40
TALC
ST. LAWRENCE CO., N.Y.
Wave Length in Microns
Wave Length in Microns


Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.41
PROCHLORITE
CHESTER, VERMONT
Wave Length in Microns
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.42
ROCK CRYSTAL
ARKANSAS
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.43
ROSE QUARTZ
CUSTER, SO. DAKOTA
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.44
ORTHOCLASE
CROOK CO., WYOMING
Wave Length in Microns


Wave Numbers in cm⁻¹
FIG.45
ORTHOCLASE VAR. ADULARIA

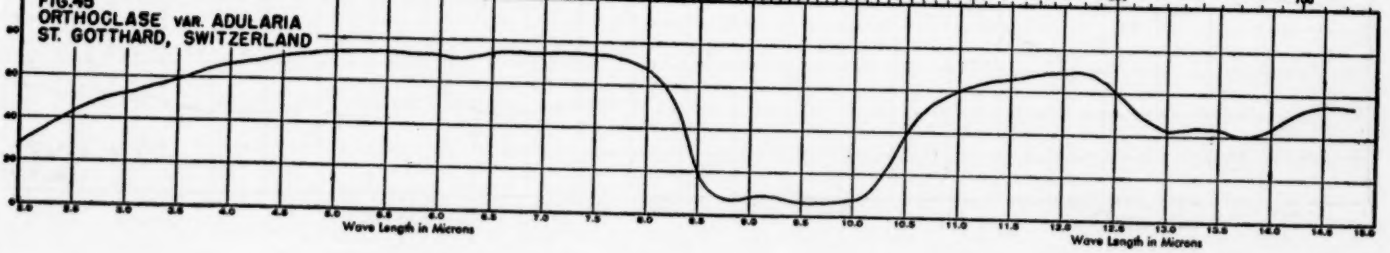
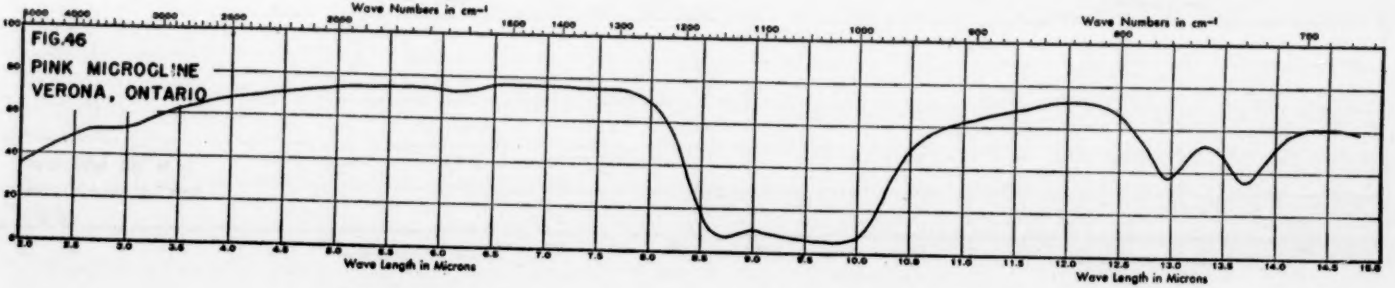
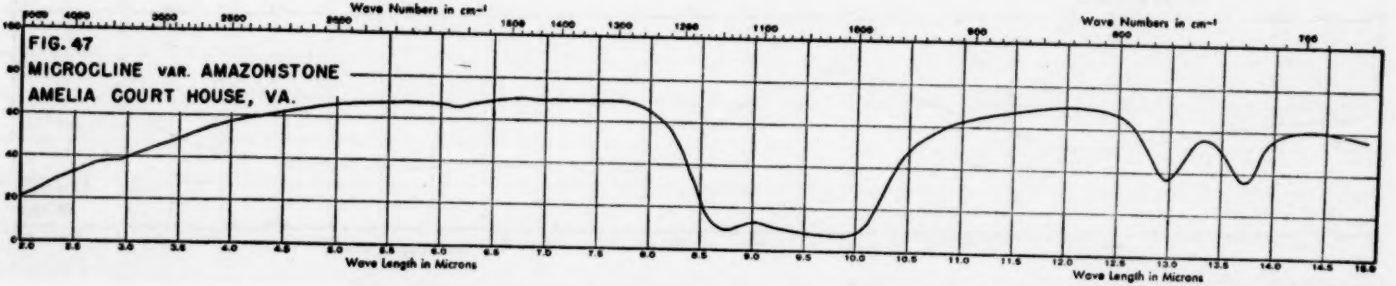
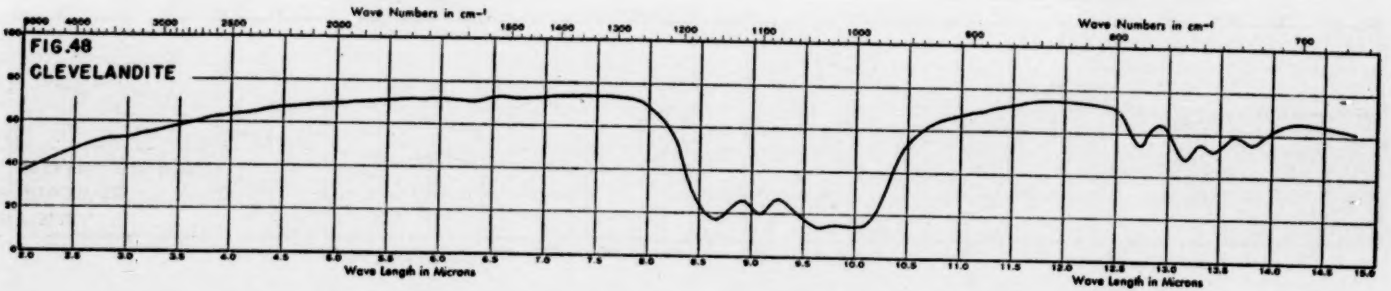
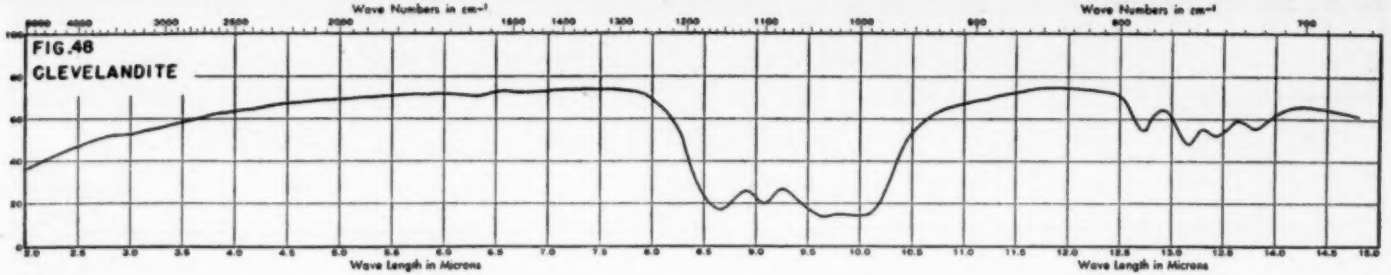


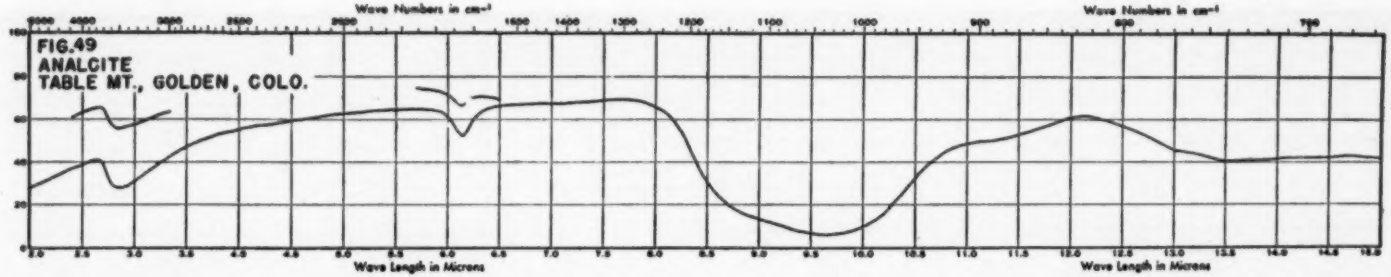
FIG.45
ORTHOCLASE VAR. ADULARIA
ST. GOTTHARD, SWITZERLAND
Wave Length in Microns
Wave Length in Microns

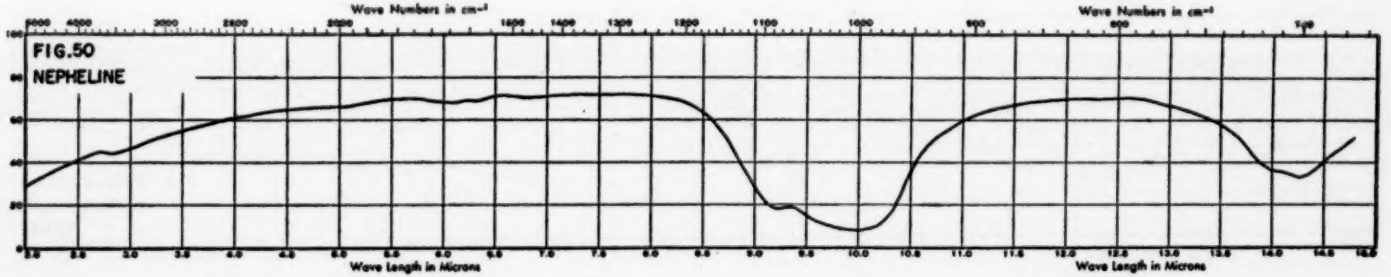

Wave Numbers in cm⁻¹
FIG.46
PINK MICROCLINE
VERONA, ONTARIO
Wave Length in Microns
Wave Numbers in cm⁻¹
Wave Length in Microns

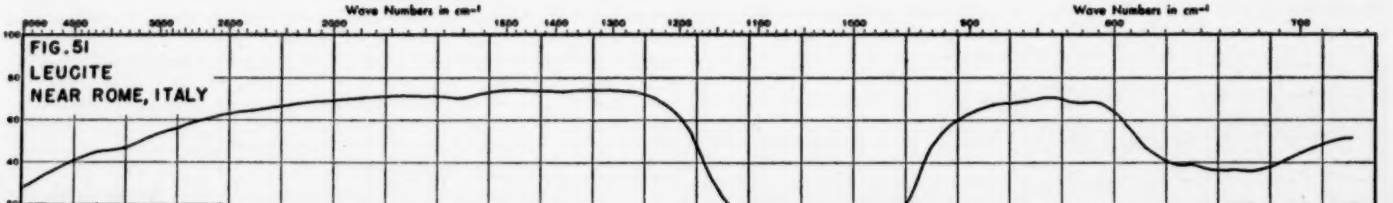

Wave Numbers in cm⁻¹
FIG. 47
MICROCLINE VAR. AMAZONSTONE
AMELIA COURT HOUSE, VA.
Wave Length in Microns
Wave Numbers in cm⁻¹
Wave Length in Microns

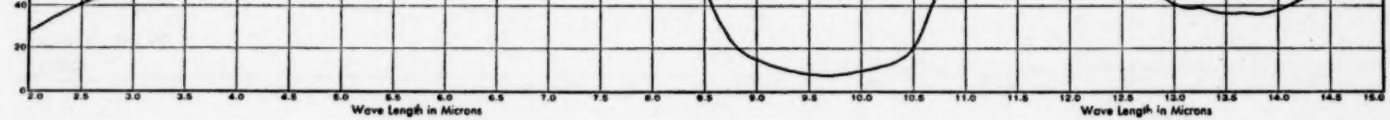

Wave Numbers in cm⁻¹
FIG.48
CLEVELANDITE
Wave Length in Microns
Wave Numbers in cm⁻¹
Wave Length in Microns

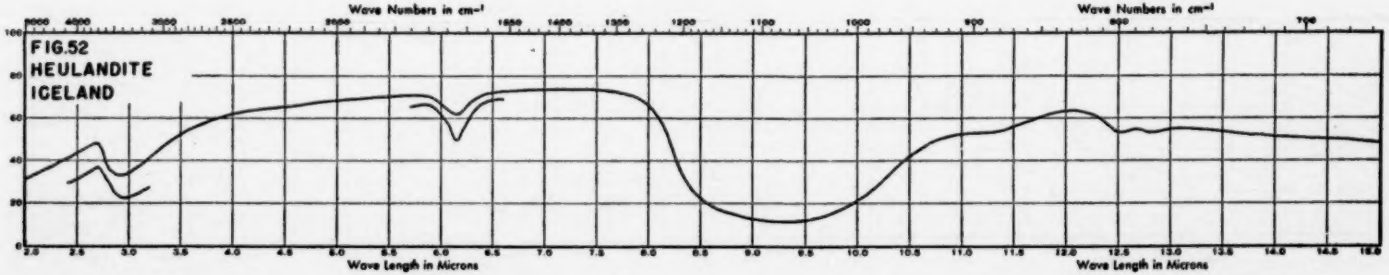

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.48
CLEVELANDITE

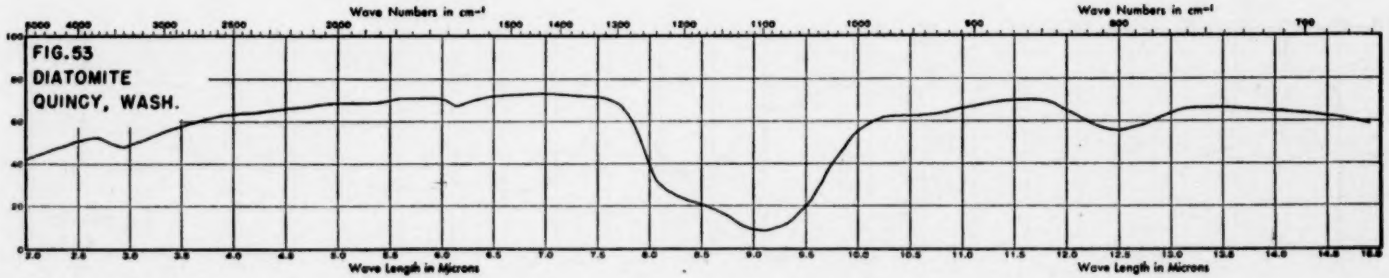

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.49
ANALCITE
TABLE MT., GOLDEN, COLO.

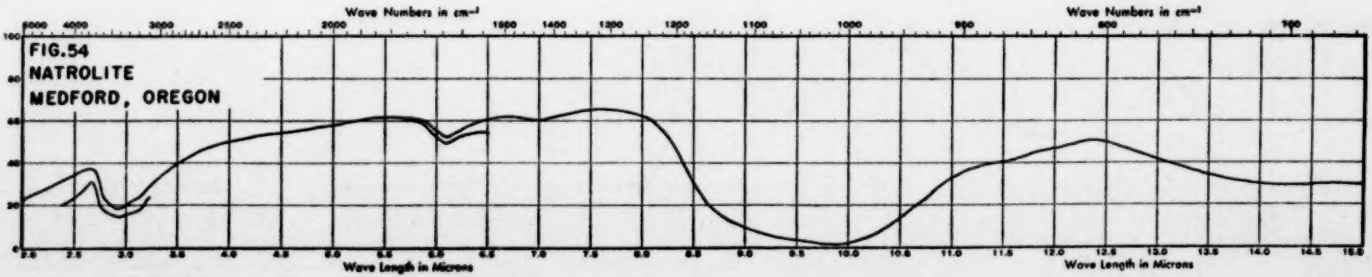

Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.50
NEPHELINE


Wave Numbers in cm⁻¹
Wave Length in Microns
FIG.51
LEUCITE
NEAR ROME, ITALY


Wave Length in Microns
Wave Length in Microns


Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.52
HEULANDITE
ICELAND
Wave Length in Microns
Wave Length in Microns


Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.53
DIATOMITE
QUINCY, WASH.
Wave Length in Microns
Wave Length in Microns


Wave Numbers in cm⁻¹
Wave Numbers in cm⁻¹
FIG.54
NATROLITE
MEDFORD, OREGON
Wave Length in Microns
Wave Length in Microns

Spectrograms of muscovite (fig. 37) and biotite (fig. 38) were obtained by using thin cleavage flakes rather than the powdered samples. The curves show better resolution in some regions than in others; the difference in resolution is attributed to the different sample thicknesses and the related scattering. The spectra have flat, low transmission areas in the regions of intense absorption because of the excessive thickness of the cleavage flakes.

Tectosilicate spectra.—The simplest (chemically) tectosilicate is quartz which was run on two specimens (figs. 42 and 43), Arkansas crystal quartz and Custer, South Dakota, rose quartz, to confirm earlier determinations (Plyler, 1929; Keller and Pickett, 1949; Adler *et al.*, 1950). The quartz spectrum is characterized by four main peaks located at 8.60, 9.20, 12.50, and 12.84 microns. Plyler (1929) assigned four fundamentals to the quartz spectrum on the basis of intensity and position; namely, 9.0, 12.5, 20.0, and 26.0 microns. On the basis of those assumed fundamentals, the other observed absorptions would be due to overtones and combinations.

The spectrum of diatomite (fig. 53), which is included to show similarities in spectral features of crystalline and noncrystalline silica, shows absorption in the OH and H_2O regions, and at 8.5, 9.10, and 12.5 microns. The main absorption peak occurs in the same region as the main band in quartz and the band at 12.5 microns corresponds in position to the doublet in the quartz spectrum.

The spectra of the potash feldspars, orthoclase, adularia, pink microcline, and amazonstone are presented in figures 44-47. The spectra of orthoclase and adularia are similar, but that of microcline differs from orthoclase in the intensity of the absorptions in the doublet region between 12.5 and 14.0 microns. Repeated runs on samples of orthoclase and microcline from other localities substantiated the differences. Hunt *et al.* (1950) also noted the sharp peaks in the microcline spectrum. The differences in the microcline-orthoclase spectra may be due to variable amounts of albite molecule present, but more probably they are due to differences in structure between the two potash feldspars like those proposed by Barth (1934). He proposed that Si and Al exist

as variate atoms in all positions in orthoclase, whereas in microcline the Al atoms are grouped in two-fold pairs.

Aluminum-containing tectosilicates were measured in the sequence where the ratio of Al to Si in tetrahedral coordination increases from 1:1 to 1:3. In the soda group, clevelandite albite, analcime, and nepheline were studied, and in the potash group, orthoclase and leucite were examined (figs. 48, 49, 50, 44, and 51). Heulandite and natrolite were included to compare with analcime (figs. 52, 54, and 49) because these are hydrated, whereas albite and nepheline are without water.

The leucite curve is simple, with broad absorption bands centered at 9.70 and 13.80 microns. It resembles the curves of orthoclase and analcime, but the absorption maxima of leucite are shifted to slightly longer wave lengths than those of orthoclase. Perhaps this shift is due to the increased Al relative to Si in leucite, which will be discussed later.

The albite curve shows notably more fine structure than other associated alumino-silicate minerals. Similar extra detail was observed by Hunt *et al.* (1950). The curves of nepheline and the zeolites show the general characteristics of the previously mentioned tectosilicates: a broad absorption band extending from about 8.0 - 11.0 microns and minor absorptions beyond 12.5 microns. Again the positions of maximum absorption are shifted slightly, but regularly, toward longer wave lengths as the Al in the Al:Si ratio is increased.

DISCUSSION OF THE MINERAL SPECTRA

Interpretation of the infrared absorption spectra of minerals still falls far short of the goal of understanding them. It has been necessary to approach the study of minerals in an empirical way, much the same as the course pursued by the chemists in their early work on organic compounds. Work on minerals is further complicated also by the fact that their bonding may vary widely between ionic and covalent types, and the extent of each type may not be well known. Indeed, the very fact that infrared absorption occurs in a mineral is evidence that its bonding is not entirely ionic in character, for the wholly ionic-bonded minerals (halite, fluorite, "artificial" AgCl) do not absorb infrared radiation in significantly measurable amounts (they absorb elsewhere). The extent of covalent bonding in minerals does not appear to have been

appreciated by mineralogists in general, judging from casual references in the literature, although Gruner (1950) and Pauling (1948) call attention to it explicitly. Suffice it to say that the intensity of absorption by the minerals varies with the extent of covalency of bonding as well as the amount of mineral in the absorption chamber. Moreover, there appears to be a mass crystal effect which also accounts for some absorption.

Despite the problems of interpretation, some progress has been made in relating the spectra to mineral structure, and as more data are assembled the opportunities and possibilities of correlating them are increased. It is reassuring to find that the positions of absorption bands for anions in minerals fall in the same general ranges as they do in chemical solutions (American Cyanamid Chart, 1950), but generally over broader ranges in the minerals. The broadening and sometimes unsharpness of mineral absorptions may possibly be due to several causes: imperfections in crystal architecture, to substitutions in the lattice, to fineness of crystals, or to incomplete crystallization, and others.

The equation relating infrared absorption to simple harmonic motion of the molecules, which has been valuable in computing the frequencies of simple bonds in chemical substances, has not been of practical help in predicting the frequencies of vibrations in complex minerals. No success has been achieved by the writers in computing "backwards" from observed data to a general case. We are still in the early stage of comparing likenesses in spectra and speculating on the correlation of groups of atoms with certain spectral properties.

The absorption spectra of quartz and silica offer an opening to more detailed consideration of the silicate spectra. Quartz is characterized by well-defined absorption within the band beginning at 8.0 to 8.5 microns and ending at 9.5 to 10 microns, lesser absorption in a double trough at about 12.3 and 13 microns, and a single one at 14.5 microns. These absorptions are found so consistently that they are thought to be diagnostic and significant of quartz. The strong absorption around 9.2 microns persists clearly also through cristobalite, opal, and fused silica (Keller and Pickett, 1949), but the absorptions at longer wave lengths may be diffuse. Per-

haps the absorptions found in quartz can be correlated, at least in kind if not in degree, to include all polymerized silica, and to clusters of silica tetrahedra.

The clearly defined, double peak absorption at about 12.5 microns in quartz is represented in other tectosilicates as a peak or as a subdued band perhaps at slightly higher wave length. It is well preserved in the 1:1 lattice clays (kaolinite group) of the phyllosilicates. Pyrophyllite, muscovite cleavage flakes, and some of the 2:1 lattice clays exhibit the 12.5 micron absorption; but in other samples of those clays it is not clearly evident. Whether the decrease in absorption is due to poor crystallinity of some montmorillonites and illites, to the fineness of the crystals, to adsorbed gels, or to unrecognized properties of those clays which may be diagnostic, is not now known.

As aluminum substitutes for silicon in the tectosilicate framework, the band of strong absorption centered about 9.0 to 9.5 microns shifts slightly, about 0.5 micron, toward longer wave lengths. Buerger (1948) observed that the mineral groups in Bowen's discontinuous reaction series (descending temperature) also run in a series of increasing number (0 to 4) of oxygen atoms shared per silicon tetrahedron. In other words, the mineral groups of Bowen's discontinuous reaction series run down temperature from nesosilicates to tectosilicates, the cyclosilicates not being mentioned. Buerger first pointed out the geometrical picture involved, and then outlined the bond picture which appeared to follow.

The infrared absorptions, which are a measure of the bonding, confirm Buerger's ideas. The position of the strongest absorption band by the silicate radical shifts slightly but regularly toward longer wave lengths from tectosilicates to nesosilicates, the cyclosilicates excepted because they vary within themselves. The high silica polymer, quartz, is apparently the closest and most tightly bonded, and therefore its absorption occurs at shorter wave lengths and higher frequencies. As the number of shared oxygens per silicon tetrahedron decreases, and as the number of aluminum and other non-silicon atoms injected into the silica assemblage increases, the position of the strong absorption band shifts toward longer wave lengths.

The absorption band in the cyclosilicates is not so closely

correlated with the silica distribution as in the other structures, but separate or isolated absorption peaks appear to characterize some cyclosilicates. Wollastonite and rhodonite, of the $A_m(B_3X_9)_n$ class, show a strong single, isolated absorption peak at about 6.9 microns, besides the band beginning in the 8.5 to 9.0 micron region. Beryl, cordierite, and tourmaline show isolated absorptions at about 8.3, 8.5, and 7.4-7.6 microns, respectively. On the other hand, benitoite does not possess an isolated absorption. The cyclosilicates may be individualistic in their absorption characteristic but an alternative explanation is that the apparently anomalous behavior may be due to improper classification of some of those minerals as cyclosilicates. They are cyclic in terms of silica tetrahedra, but from the viewpoint and classification of Strunz (1937) (see also Gruner, 1948, p. 679) other tetrahedral configurations may be of first order significance. The cyclosilicates deserve more detailed attention.

If, as Plyler suggests, the fundamental frequencies of silicate absorption fall predominantly at longer wave lengths than 15 microns, supplementary measurements by other instrumentation than that which is at hand might open up means of computing the relationships between all silicate infrared spectra and structure in a more detailed manner.

ACKNOWLEDGMENTS

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REVIEWS

Middle Cambrian Stratigraphy and Faunas of the Canadian Rocky Mountains; by FRANCO RASETTI. Smithsonian Miscellaneous Collections, vol. 116, no. 5. Pp. v, 277; 34 plates, 5 figs. Washington, 1951.—The term Albertan has long been employed in texts as the name of the Middle Cambrian series. However, researchers have consistently used Appalachian or Montana sections as a standard faunal sequence for these strata, largely because Walcott's early pioneer work in the Canadian Rockies has been generally recognized as provisional and sketchy. The work here reviewed may well change this situation and is therefore an important contribution to Cambrian biostratigraphy.

An interesting introduction and a summary of previous work are followed by 44 pages of measured sections with comments. The area studied follows the Canadian Pacific Railroad from south of Lake Louise, Alberta, to Field, British Columbia, including peaks lying within 5 or 10 miles north and south of the railway. A section on stratigraphy redescribes 2000 to 3000 feet of beds divided into the Lower Cambrian St. Piran sandstone (with an upper Peyto limestone member) and the Middle Cambrian Mount Whyte dark shale and limestone, Cathedral dolomite, and the Stephen dark shale and limestone. The Mount Whyte and Stephen formations include some local developments of fossiliferous shale. The Lower and Middle Cambrian trilobite faunas of these beds are discussed in sequence, and a zonation tentatively offered at least for the Cordilleran geosyncline. Over 120 pages are devoted to systematic descriptions of trilobites followed by 26 plates of fossil illustrations. As in the author's previous works, the photography is superior, the material being carefully prepared and coated. Six new trilobite genera and about 70 new species are described.

Certain parts of the work are considered particularly noteworthy for Cambrian stratigraphers and paleontologists. (1) The stratigraphy confirms Walcott's delineation of at least two Cambrian sedimentary troughs in this part of the northern Cordilleran geosyncline. Mount Stephen, British Columbia, lies athwart these. The Bow trough to the east contains a carbonate algal reef facies of the Cathedral dolomite. This grades laterally westward along Mount Stephen into dolomite tongues within shale and limestone like that of the Stephen formation above. The Cathedral interval thickens westward from 1000 to over 2000 feet, and the Mount Whyte formation below thickens greatly in the Mount Stephen-Mount Field area. Rasetti considers this location to mark the boundary between the easterly Bow trough and the westerly Goodsir trough, Middle Cambrian strata here already showing the difference in facies made known in Walcott's early work on the Upper Cambrian.

(2) The stratigraphic positions of the Burgess and *Ogygopsis* shale lentils are discussed in descriptions of the Mount Stephen and Burgess Quarry sections. The famous faunas occur in different fault blocks, and accurate field determination either of their relative position or placement in a normal section is impossible. Rasetti on paleontological grounds correlates the *Ogygopsis* shale on Mount Stephen with the lowest exposed faunas of the Stephen formation in the Burgess Quarry block across the Kicking Horse River from Mount Stephen. Walcott's remarkable "Phyllopod bed" of the Burgess shale lies toward the top of this faunal succession. The Burgess is seemingly a local shale development and not represented in normal sections of the Stephen. Its fauna is judged older than Rasetti's highest Middle Cambrian fauna (*Bathyuriscus-Elrathina* zone) but younger than the *Glossopleura* zone. It falls in the middle shaly part of the Stephen formation and in the *Ogygopsis* shale which is evidently a little lower. Walcott considered the latter to be in the upper Stephen, but Rasetti's faunal evidence for the lower correlation is convincing.

(3) Paleontologic contributions of the paper exceed the usual describing of new trilobite genera and species. The author reviews his previously published ideas on trilobite classification (with which this writer agrees) and emphasizes the division of opisthoparian trilobites into two superfamilies. The former name of one of these, Bathyuriscidea, has been replaced for priority reasons with the name Corynexochoidea. The Zacanthoididae are placed in this group along with the more typical bathyuriscids. Attention is again called to the lack of logical subdivisions within the other superfamily, the Ptychoparoidea, which now includes over 75 Lower and Middle Cambrian genera from North America and several hundred Upper Cambrian and Lower Ordovician forms.

(4) In discussion of the ptychoparoids Rasetti criticizes the use of special assigned generic and specific characteristics in classification. The criticism seems just to this writer. Certain phylogenies of ptychoparoids are becoming fairly obvious both in Middle and Upper Cambrian trilobites. Study of these will probably indicate that characters stable enough in one lineage to be considered generic will show great variability within other phyla. For example, Rasetti has stated that the width-of-fixed-cheeks ratio, a "generic character," may vary as greatly within certain ptychoparoids as the brim-border ratio, a "specific character," does in other groups with which this writer is familiar. A clear designation of certain traits as generic and specific may well lead to an artificial classification rather than to a natural one.

(5) The author's restriction of the Lower Cambrian to the range of olenellid trilobites may yet be open to controversy when

work similar to Rasetti's has been done elsewhere, but the procedure seems entirely logical in this area. The highest Lower Cambrian zone, *Kochiella-Inglefieldia-Syspacephalus* of Resser and Howell et al. (1944) seems best abandoned, being overlapped and superseded by the *Bonnia-Olenellus* zone in the Rocky Mountains.

(6) Rasetti inserts two new Middle Cambrian faunal units in his Albertan sequence. These, the *Wenkchemnia-Stephenaspis* and *Plagiura-Kochaspis* zones fall beneath the *Albertella* and *Glossopleura* zones. The higher Middle Cambrian zones of Deiss (1938) and Howell et al. (1944) are not accepted by Rasetti. No clear subdivision of the post-*Glossopleura* faunas based on the Montana zones of Deiss are possible in Alberta. These faunas are grouped by Rasetti into the *Bathyuriscus-Elrathina* zone, which is subdivided locally, but as a unit is sharply distinct from the underlying *Glossopleura* zone, indicating a faunal hiatus within the Stephen formation. Rasetti is careful to distinguish local and possibly facies-controlled faunules from the larger units which may form the basis for a future frame of reference for continent-wide correlation. This fact, plus the excellence of the author's illustrations and taxonomic descriptions, should encourage all Cambrian researchers to turn again to the southern Canadian Rockies for a standard Middle Cambrian section. JAMES L. WILSON

The Tectonics of Middle North America; by PHILIP B. KING. Pp. xix, 203; 52 figs. and frontispiece. Princeton, 1951 (Princeton University Press, $3.75).—Tectonic geology, which deals with the larger units of geologic structure and their genetic relationships, has developed remarkably since the turn of the century. Literature in this field is voluminous, grows at an accelerating pace, and becomes obsolete in varying degree. Practising geologists as well as students welcome a work that undertakes to present the current status of the subject for any large geographic unit, with some appraisal of the pertinent literature. King's treatise, "An outline of the structural geology of the united States," issued in 1933 as a guidebook of the Sixteenth International Geological Congress, was a first attempt to perform such a service for this country. With the advantage of more mature knowledge and judgment, and on a far broader base of published information, King is now preparing a more extensive structural analysis of the entire United States and contiguous areas in Canada and Mexico. The lithoprinted volume here reviewed considers all of the territory east of the Cordilleran front. A complete edition to appear later will include the broad and complex Cordilleran belt.

Analysis starts with the central stable region, made up of the Laurentian Shield and the wide interior lowlands that border the

shield on the south and west. Two chapters devoted to these major units make up about a third of the text. Half of the total space is given to "Paleozoic structures southeast and south of the central stable region"; that is, to the Appalachian, Wichita, and Ouachita mountain systems. Finally the Coastal Plains are treated in two brief chapters. Each of nine chapters in the book has a selected bibliography, presumably made up of source materials that the author found most helpful, and recommended by him as supplementary reading. Sources of the 52 text figures also are listed. Study of the book is facilitated by an adequate index as well as a detailed table of contents.

Critical readers will find the treatment well balanced, and will feel that the author has combed the source materials with care, keeping of the older concepts only what has stood the test of time and using effectively all important new information. The student's time is saved by reduction of explanatory text to a minimum, with excellent text figures to provide an effective supplement. Perhaps condensation is somewhat overdone; readers would like to know, for example, the evidence for statements such as the following on a part of Newfoundland (p. 88): "... the southern crystalline rocks are probably of Paleozoic age and are not basement (as they are represented on the Tectonic Map of Canada)." The overall picture of the Appalachians is much sharpened by results of new studies, particularly in Newfoundland, New England, and the southern part of the belt. Growing evidence that early- and mid-Paleozoic orogeny extended the full length of the chain gives a consistency that was lacking in the older views of Appalachian development.

In a brief chapter introductory to his account of the Appalachians King presents "general and theoretical considerations" to help the reader integrate the complex details of structure. Evidence cited by earlier workers for a two-sided or wedge-shaped cross-section of the mountain belt is amplified, and the concept of a tectogene formed by sharp downbending of a sialic shell is championed. This assumed mechanism is open to serious question. The diagram (fig. 16) copied from Hess' representation of the Alps is badly out of scale, since with the assumed thickness of sial (25 km.) the width of the mountain chain would be no more than 60 km.; actually it is 125 to 175 km. Moreover, in a typical island arc, such as the East or West Indies, the normal thickness of sial probably is much less than 25 km. Regardless of the thickness, failure by downbuckling seems extremely improbable. Kuenen's laboratory experiments produced such buckles in a layer with appreciable strength resting on a substance devoid of strength. This relationship does not accord with our knowledge of the earth's crust. Sial is

not inherently stronger than sima, and since mafic rocks have a much higher melting point than silicic rocks, higher temperature at moderate depths can not be invoked to give the desired difference in strength at a sharply defined level. A mountain "root" may be formed by plastic deformation, but probably not in the way suggested by King's figure 16.

Diagrams in the book are uniformly excellent. Sections across belts of strongly deformed strata have vertical and horizontal scales the same, thus avoiding the distortion that mars so many geologic structure sections. On the other hand some vertical exaggeration is desirable in long sections across lowlands beneath which stratified formations are but mildly deformed. All such sections in King's book have uniform scales, with the vertical ten times the horizontal, and thus readers can compare them directly, without effort.

The tectonic patterns and history of the Cordilleran region present a large challenge. Readers of King's preliminary volume will await with interest the complete edition.

CHESTER R. LONGWELL

The Dynamics of Faulting and Dyke Formation, With Applications to Britain, 2d ed.; by E. M. ANDERSON. Pp. x, 206; 39 figs. Edinburgh and London, 1951 (Oliver and Boyd, 22s. 6d.).—Close comparison of this edition with the original, which appeared in 1942, reveals only small changes, chiefly additions, with a net increase of 16 pages. Nevertheless structural geologists should be notified that this important book is again available. Anderson's analysis of faulting makes use of mechanical principles that an engineer can endorse. His opening sentence merits quotation, with emphasis: "In dealing with the dynamics of faulting the terms pressure, tension, and shearing or tangential stress will be in constant use." He goes on to point out the common error of regarding pressure and tension as *forces;* his own treatment deals, as it should, with stresses and related strains. At the outset he leads the student into three-dimensional geometrical concepts, the *sine qua non* in structural geology. Moreover he introduces in early pages, and emphasizes repeatedly thereafter, the basic principle announced in 1833 by Navier, which explains that shearing stress is at a maximum along planes at 45° to the greatest pressure, but that planes of shearing rupture make a smaller angle to the axis of compression. No reference is made to the strain ellipsoid concept, the use of which by some textbooks in connection with shearing rupture throws students into confusion from which they are rescued only with much difficulty.

With all of its excellence, the book has some curious weaknesses. Geologists who have studied the prominent breccias associated with extensive thrusts in Nevada and California will read with astonishment Anderson's pronouncement (p. 19), "It has been observed that fault-breccias frequently occur along normal and wrench faults, but not in association with thrust-planes." He goes on to explain that this is in accord with dynamical theory. But surely a theory that is out of accord with clear field evidence is in need of repair. Lack of breccias along thrust surfaces known in Britain will not establish Anderson's thesis, since a mechanical principle must have general application. The apparent contradiction between evidence in Britain and that cited from southwestern United States appears to have a logical explanation. Some of the extensive plates preserved in the latter region were propelled across irregular land surfaces, and coarse breccias were developed, perhaps under moderate load. On the other hand the thrusts exposed in the Scottish Highlands probably evolved at great depth; the surfaces are extensively lined with mylonite, which is formed only in movement under extreme pressure, according to an able analysis by Louderback (Seismol. Soc. America Bull., vol. 32, 1942, pp. 305-330). Thus Anderson's rule regarding fault breccias requires important modification.

We read also (p. 20): "It follows from the dynamical theory that thrust planes and normal faults can never occur in association." Again the theory fails the test of clear field evidence. West of Las Vegas, Nevada, the trace of the Keystone thrust changes direction from north to N. 50 E. as it approaches a zone of strike-slip faulting. In this area of complex movement a part of the thrust plate, apparently under torsional strain, failed on several large normal faults, and the resulting blocks were in part overridden as the unbroken portion of the plate continued to advance (see W. S. Glock, Am. Jour. Sci., 5th ser., vol. 17, 1929, pp. 326-341). Probably torsional strain has not been uncommon during thrusting, particularly in the movement of plates over irregular land surfaces. All faults record failure under shearing stress, and Anderson's concept of normal faults as generically unique does not seem reasonable. His view that all steep reverse faults fall basically into the wrench (strike-slip) class also is unconvincing.

The numerous British examples of faults and fault systems discussed in the book are highly instructive. CHESTER R. LONGWELL

Tables for Microscopic Identification of Ore Minerals; by W. Uytenbogaardt. Pp. vii, 242. Princeton, New Jersey, 1951 (Princeton University Press, $5.00).—The purpose of this book is to

provide a set of determinative tables for the identification of ore minerals when observed in polished surfaces under the microscope. It includes all adequately described ore minerals up to 1950.

The book is designed as a laboratory handbook and the characteristic used as a primary diagnostic criterion is the polishing hardness, as manifested by the relief of the different minerals on the polished surface. The minerals are grouped in the tables according to their hardness as compared with galena, chalcopyrite, or pyrite, these three minerals being chosen because they are common in many ore associations and are easily recognized. The data for each mineral are arranged in the following sequence: name, chemical formula, crystal system, Talmage hardness, reflectivity, color, etch tests, and miscellaneous (internal reflection, twinning, cleavage, intergrowths, associated minerals, and other features). The author emphasizes that although numerical values of reflectivity require photometric equipment not often available for routine work, qualitative estimates are diagnostically useful, since the differences between individual minerals are more significant than the absolute values. The color is given as seen in oil immersion, and additional information is provided for the appearance against the usual associated minerals.

In addition to the tables themselves, two useful lists are included; one gives the ore minerals in order of increasing polishing hardness, the other in order of increasing reflectivity. An appendix enumerates superfluous mineral names, with a brief account of each one and the relevant references. The book concludes with a most valuable bibliography of 441 separate articles, representing a comprehensive cover of the literature of ore mineralogy over the last twenty years.

The book appears to be agreeably free from misprints and errors. It is well set out and printed by the photo-offset process.

The originality of this work lies in its emphasis on physical properties, particularly the easily recognized polishing hardness, as diagnostic criteria for ore minerals in polished surfaces. The author gives the characteristic etch reactions, but as secondary means of identification, and in the introduction he warns against the ambiguous results such tests are liable to give.

In the opinion of the reviewer, this book is not only an admirable guide to the ready identification of ore minerals but also an excellent compilation of the best data on their properties, associations, and paragenesis. It contains an enormous amount of information in a brief compass. Its completeness and up-to-dateness and the reasonable price should commend it to mineralogists and economic geologists.

BRIAN MASON

Die Lagerstätte der Trepca und ihre Umgebung; by F. SCHUMACHER. Pp. 65; 18 figs., 2 colored plates. Belgrade, 1950 (Saveta Za Energetiku I Ekstraktivnu Industriju Vlade Fnrj, 150 Jugoslavian dinars).—In this publication Professor Schumacher describes in some detail the Trepca lead-zinc deposit in southern Serbia, where the principal ore body is associated with a volcanic plug at the intersection of two faults.

The Stari Trg district, in which the Trepca deposit is situated, is underlain by a folded metamorphic complex of early Paleozoic slates, quartzites, and limestones. The folding is related to early Tertiary Dinarian orogeny. This folded basement is overlain by extensive Miocene lavas and tuffs.

The principal fault in southern Serbia extends from Saliniki northwesterly to Bosnia. A branching fault at Mitrovica extends to Belgrade. Lead-zinc deposits of late Miocene age lie at several localities in the faults, the principal deposit being at the intersection of the two faults. At this locality a volcanic plug was intruded along an anticlinal core of limestone, where the crest of the anticline had been fractured by prior folding. The volcanic plug is described as an explosion breccia with a trachytic core.

The ore body lies along the footwall of the volcanic chamber and fingers into the limestone. The average ore area for five levels is 88,200 sq. ft., and the average grade mined to date is approximately 8.8 per cent lead, 6.0 per cent zinc, and 3.7 ounces per ton silver. There has been a progressive lowering of grade, particularly zinc, with depth.

Professor Schumacher describes in detail the contact metamorphic effects of the volcanic intrusion on the limestone with the development of skarn silicates followed by hydrothermal alteration of both the volcanic breccia and limestone. He ascribes the deposit to a contact metamorphic origin with a predominantly hydrothermal paragenesis, especially emphasizing the combination of pyrometasomatism and hydrothermal metasomatism in a falling temperature sequence.

Professor Schumacher does not hold much hope for the discovery of other large ore bodies in the district, as he considers the Trepca deposit the result of an unusual combination of geological conditions.

J. D. BATEMAN

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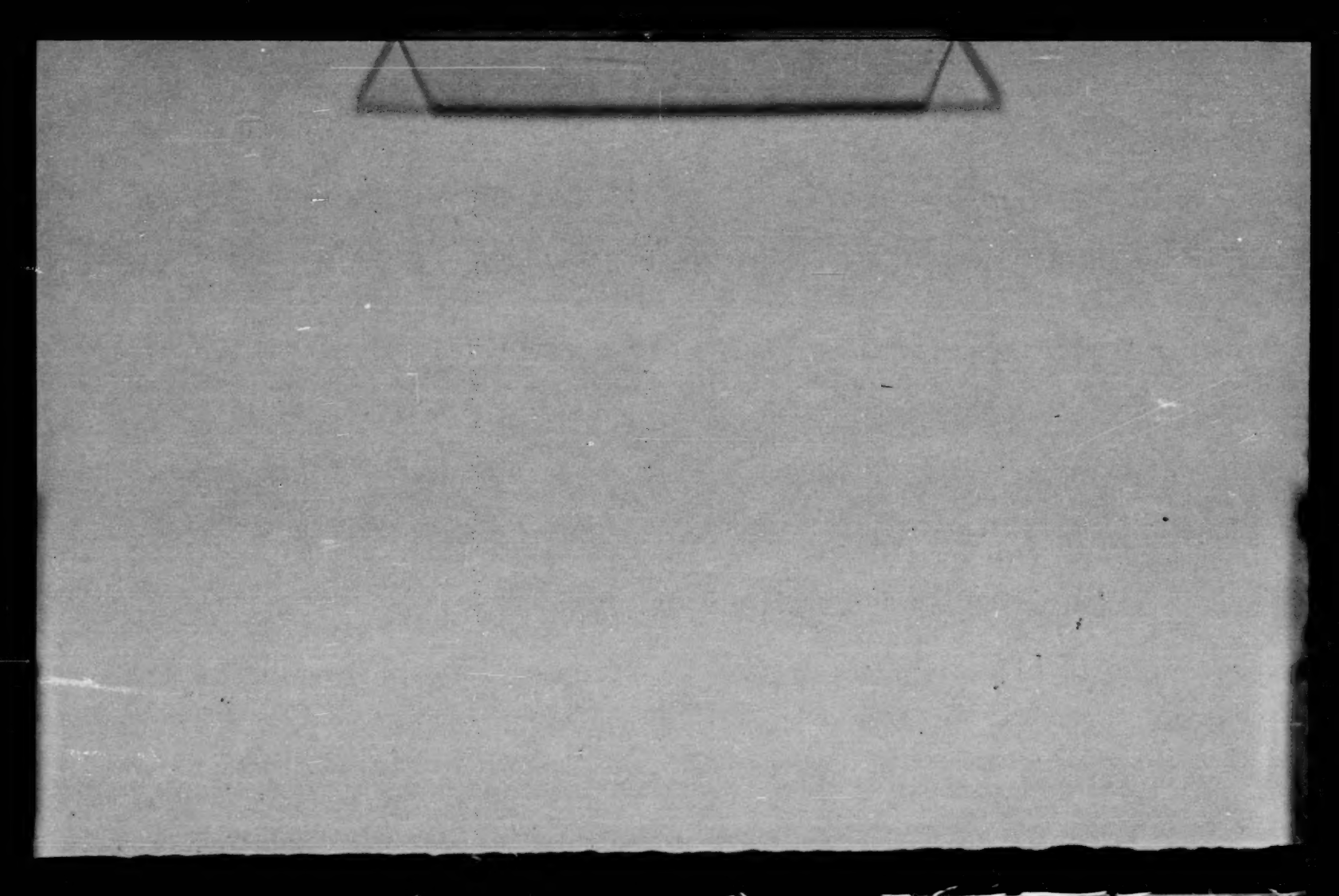

VOL. 250 No. 6, 1952

CONTENTS

PLIOCENE-PLEISTOCENE BOUNDARY IN CALIFORNIA COAST RANGES ..... *W. P. Woodring* 401

ABSOLUTE AGES OF RADIOACTIVE MINERALS FROM THE APPALACHIAN REGION ..... *John Rodgers* 411

A NEW QUANTITATIVE METHOD OF TOPOGRAPHIC ANALYSIS *Carleton A. Chapman* 428

INFRARED SPECTRA OF SOME ROCK-FORMING MINERALS *W. D. Keller, John H. Spotts and Donald L. Biggs* 453

REVIEWS

Middle Cambrian Stratigraphy and Faunas of the Canadian Rocky Mountains; by FRANCO RASETTI ..... *James L. Wilson* 472

The Tectonics of Middle North America; by PHILIP B. KING *Chester R. Longwell* 474

The Dynamics of Faulting and Dyke Formation, With Applications to Britain, 2d ed.; by E. M. ANDERSON ..... *Chester R. Longwell* 476

Tables for Microscopic Identification of Ore Minerals; by W. UYTENBOGAARDT ..... *Brian Mason* 477

Die Lagerstätte der Trepca und ihre Umgebung; by F. SCHUMACHER ..... *J. D. Bateman* 479

PUBLICATIONS RECENTLY RECEIVED 480

PAYNE & LANE, *Printers* — NEW HAVEN, CONN.